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GEOLOGICAL GEOMORPHOLOGY

BY RICHARD J. RUSSELL

(Address as Retiring President of The Geological Society of America)

ABSTRACT

Geomorphology has not developed as substantially as Gilbert forecast in 1890. Physiography concentrated on problems of erosion, almost to the exclusion of other parts of the discipline, and developed a terminology which became elaborated beyond usefulness. Disregard of the third dimension and inadequate geophysical backgrounds led to unrealistic results by physiographers.

This address is concerned primarily with alluvial morphology, a field neglected or commonly misinterpreted by physiographers. Flood-plain deposits, deltas, and deltaic plains are considered with reference to the sedimentary, structural, and morphological processes under which they originated, as examples to illustrate the value of a more geological geomorphology.

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INTRODUCTION[1]

"When the work of the geologist is finished and his final comprehensive report written, the longest and most comprehensive chapter will be on the latest and shortest of the geological periods." When writing these words, Gilbert (1890, p. 1) believed that intensive study of the Quaternary would reveal many of the keys most useful for the interpretation of earth history.

But in spite of the advantages offered by this short period, the ease of deciphering its record, an abundance of evidence, its readily understood environmental conditions, and the opportunity of observing processes occurring in the laboratory of nature, geologists have not concentrated on the final chapter of the comprehensive report to the degree forecast. Quaternary studies gradually lost an aura of respectability which is attached to "hard-rock" geology.

That geomorphology has not attained the stature which appeared probable prior to the turn of the century in part appears to be related to the rise and fall of physiography. That discipline gained prestige during the scientific

[1] Anatolian, Brazilian, and some other portions of the text are based on field work conducted under Project No. N7onr35608, Task Order No. 388 002, Geography Branch, Office of Naval Research. Text figures were prepared under the direction of Philip B. Larimore.

exploration of the West but later lost ground because its practitioners were prone to solve problems which they were ill prepared to tackle.

PHYSIOGRAPHY

Historical Summary

Physiography was originally the broad science of nature, with description emphasized, rather than theory or explanations. Under William Morris Davis the description became explanatory, and the field narrowed to physical geography.

When Gilbert, King, Powell, I. C. Russell, and other students of surface evolution during the latter half of the nineteenth century were discovering fundamental principles of landform development, there was no question as to whether their work was geological. But to the extent that physiography flourished during the earlier decades of the present century, there was a corresponding tendency to relegate the field to Geography. It was entered by a large number of persons who lacked requisite training in geophysics and geology.

The concept of the erosion cycle accounted for a rush of peneplain hunters who were likely to regard anything from alleged accordance of summit levels to broad alluvial flats as evidence of a completed cycle. The quest eventually lost popularity, however, so that within recent years the rate of peneplain discovery has come to a near halt.

Peneplain enthusiasts commonly disregarded the effects of isostatic compensation. They failed to recognize the proposition that reduction in actual relief is many times slower than the rate of denudation of a rising rock column, so that sharply crested ridges and deep valleys persist for enormous intervals of time. Nor did they give adequate thought to the possibility that, by the time flatness develops worthy of recognition as a peneplain, not only the sedimentary sections of mountains but also much of their rising rock columns will have been denuded. In its penultimate condition abroad denuded region will be shieldlike, and the presence of former mountains may be determined by evidence such as the attitudes of belts of crystalline and metamorphic rocks, lineations, and fracture patterns. It is doubtful whether widespread parts of the crust which reach this condition experience further diastrophic activation. If they do, it is probable that the stage must be prepared by a long period of sedimentary-rock accumulation.

Many of the structural interpretations of the physiographer failed to be useful or, indeed, sound. Geologic history is clear as to the fact that certain parts of the crust tend to act as positive areas, while others behave negatively, whenever afforded a diastrophic opportunity. Formations thin, and parts of sections tend to be absent in the positive areas, whereas all stratigraphic units tend to thicken in structural basins or troughs. Conclusions based on landforms which postulate numerous reversals between uplift and depression are likely to prove invalid.

In the days of Gilbert and Powell it was useful to invent a reasonable amount of scientific terminology for the description and interpretation of landforms, but during the rise of physiography the vocabulary became so detailed and specialized that the subject degenerated into a semantic game. To conclude that a stream was consequent but had an obsequent extension became a research end in itself. A paper might be regarded as incomplete if it failed to introduce new terms for concepts which had earlier been explainable concisely, in English.

Some terms in the jargon are vicious, for example, "pediment." I may be ignorant of the fact that an author has quite a different understanding from my own while reading his paper. I have been amazed in the field to discover the variety of things that are lumped under this term. Some nouns have acquired entirely new meanings; for example, Andersson's (1906) thoroughly adequate definition of solifluction became an entirely different process under Eakin (1916, p. 76). This confusion, of course, is not monopolized by physiographers. No better example could be cited than what happened to Dana's (1873, p. 432) "synclinorium," which, in spite of being one of the most thoroughly explained definitions in geological literature, changed from mountain to structure in the hands of Van Hise (1896).

Hampered by elaborate terminology and growing more detached from evidence acquired by other disciplines, physiography declined. Its chief survival is in Europe, and especially in the "pure morphology" of the British. Students of landforms, in breaking away from physiography, commonly adopted the label geomorphologist. More important, with the change has come closer association with related fields such as structural geology, petrology, sedi-

mentation, stratigraphy, paleontology, and pedology.

Concern with Erosion

The physiographer concentrated on problems of erosion, but he tended to restrict his observations to the surface. He exhibited remarkably little curiosity either about the third dimension of alluvial deposits or the consequences of the erosion he postulated.

Flat flood plains were ordinarily regarded as erosional features. Fenneman's (1906) *Flood Plains produced without floods* standardized a concept which persists in our modern textbooks. The surface of the broad alluvial valley of the Lower Mississippi River has been regarded as the result of spreading a thin veneer of alluvium across a bedrock flat which was produced by lateral corrasion. In actuality the alluvium is much thicker than the deepest scour pools along the river, and it deeply blankets a highly irregular pre-Recent oxidized surface which is known in considerable detail (Fisk, 1944, Pl. 3).

The emphasis on erosion stems from Davis (1899, p. 495; 1909, p. 267), whose numerous publications ordinarily neglected processes involved in mass movement and deposition. His great concern with the erosional activities of rivers is expressed in the statement, "Ordinarily treated, the river is like the veins of a leaf; broadly reviewed, it is the entire leaf." The mechanism for erosion was thus extended "all over the basin up to the divides." Though Davis and his followers recognized the possibility of depositional terraces, in practice they attributed practically all flats to lateral corrasion, regardless of the thickness of alluvium beneath them. Even the extensive coastal flats around the Gulf of Mexico were attributed to wave planation, in spite of obvious origin as depositional deltaic plains which today preserve initial surface details such as systems of Pleistocene natural levees (LeBlanc and Bernard, 1954).

The erosion concept was extended to the continental shelves, and for their planation a theoretical wave base was set at a depth of 100 fathoms (Gulliver, 1899; Johnson, 1919 p. 80–83). While currents capable of scouring and transporting sediments occur at extreme depths in oceans (Dietz and Menard, 1951; Menard, 1952; Ericson, Ewing, Heezen, and Wollin, 1955), and fine materials are winnowed actively from bottoms such as the rocky shelf of Norway (Holtedahl, 1955, p. 130–134), there appears to be overwhelming evidence against erosional origin for continental shelves generally. Overlooked by the physiographers who postulated the 100-fathom depth for wave erosion were the conspicuous changes of level of sea and land which characterized the last chapter of geological history. The rise of the seas during the Recent waning of continental ice has been at least 450 feet (Fisk, 1944; 1955; Fisk and McFarlan, 1955; Russell, 1948; 1958).

The 100-fathom depth was an unhappy choice. The shelves ordinarily lie closer to half that depth. But the limit of effective wave erosion is conspicuously less, commonly about 6 fathoms. Borings along the Outer Banks of the mid-Atlantic coast and on linear islands flanking the Gulf of Mexico usually reveal typical beach deposits to a maximum depth of 30 feet or so. In some of the pronounced "break throughs" in the vicinity of Cape Hatteras they extend to slightly more than twice that depth. Beach deposits overlie a sharp break, below which are sediments of other origins, such as Pleistocene marsh silts, which are being covered by the encroaching beach. Nowhere on the Louisiana shelf or continental slope is there a flat caused by wave erosion, at 100 fathoms or other similar depth.

The depth of the outer margin of a continental shelf is unrelated to wave erosion. Nor is it a good criterion for estimating the level of pre-Recent seas. Shelves have been subjected to warping, faulting, and, most importantly, to shoaling by the accumulation of blankets of Recent deposits such as sand, silt, volcanic debris, and organic limestone. In most cases the level of pre-Recent, Pleistocene, low-stage seas was considerably lower than the existing shelves.

The contact between the base of the Recent and the materials it overlies is sharp on the Louisiana shelf. Above it the section consists of chemically reduced sediments, commonly with basal gravels. Immediately below the contact there is an oxidized surface which was formed during the last low-sea stand. Twenty miles south of the mouth of the Mississippi this contact occurs at a depth of 850 feet, and toward the edge of the shelf at 900 feet. There is no reason to believe that the thickest section of Recent materials has been discovered (Fisk and McFarlan, 1955, p. 286), but it is apparent that a significant lens occurs in the area where deposition was heaviest during the past few thousand years. To the east and west of the Lower Delta the base of the Recent commonly occurs at about 550 feet, and the shelf depth, at

the top of the Recent, is as shallow as 50 feet some 40 miles offshore in central Louisiana.

The position of the base of the Recent is not the best guide for determining the level of pre-Recent Gulf of Mexico. Less disturbed than the shelf is the gradient of the trench of the Lower Mississippi River, which is quite uniform from Cape Girardeau Missouri, to the Louisiana coast. At Houma, nearly 50 miles inland, the trench is somewhat deeper than 400 feet. If this gradient is extrapolated across the shelf to the shore of the time, a Gulf level of about −450 feet is indicated. Thus the observed depth of 550 feet for the Recent-Pleistocene contact appears to indicate about 100 feet of Recent subsidence. Eastward, below the delta, the subsidence is much greater.

This evidence from the most thoroughly explored continental shelf on earth refutes the physiographer's lateral-corrasion hypothesis for the Lower Mississippi Valley and an erosional explanation for the shelf. A deeply alluviated major valley leads to a deeply blanketed shelf. The flatness is depositional in both cases.

Consequences of Erosion

The physiographer paid scant attention to the consequences of erosion. In desert basins, where the case is relatively simple and the end product is obvious, he made little attempt to view the problem in quantitative terms. Rather than estimate the quantity of rock which must have been removed from mountains in order to fill near-by basins and produce today's landscapes, the physiographer took the course of trying to explain inselbergs, widespread alluvial aprons, and various basin flats on the basis of erosion and the lateral shifting of streams. He regarded alluvial deposits as veneers in spite of evidence from occasional borings which appeared to verify Lawson's (1915) "epigene-profiles" thesis or, at least, indicated profound amounts of fill.

Failure to think in quantitative terms extended beyond physiographers or desert basins. It required the rise of petroleum geology to advance paleogeography beyond elementary exercises in finding shore lines and deducing directions whence sediments came. With studies based on numerous borings came realization that extremely thick sections of sedimentary materials occur in basins of limited area. We on the Gulf Coast are directly concerned with the quantitative aspects of such problems. The consequences of erosion strike us forcefully when we find some 300,000 cubic miles of Mesozoic and Cenozoic deposits in the central Gulf Coast region alone (Murray, 1952), or at least 8000 cubic miles of Late Quaternary sediments concentrated in the vicinity of the Lower Mississippi River Delta (Fisk and McFarlan, 1955).

As an interesting sidelight we may note that the parts of the northern Gulf Coast section that have been subjected to the most elaborate differentiation in terms of series, group, formation, or zone names are the parts with the least depositional, structural, or volumetric significance. A short interval characterized by some marine deposition, but during which a trivial amount of rock accumulated, may exhibit several contrasting faunas and considerable lithologic diversity. An excellent example is the Jackson-Vicksburg section, with some 500 feet exposed along the outcrop and maximum subsurface thickness of possibly 2500 feet, which has been divided into at least 50 named units. In contrast, the Midway-Wilcox, with a thickness approximating 6000 feet, attracts comparatively little nomenclatural attention. It was deposited mainly in fresh or brackish waters, and its fossils are generally few and uninteresting. But from the standpoint of volume the Midway-Wilcox represents one of the major "waves of alluviation" that has affected the northern Gulf Coast since Paleozoic time. Its known volume in this area is on the order of 50,000 cubic miles, and the presence of this huge mass of rock accounts for highly significant structural complications in the generally homoclinal Mesozoic-Cenozoic section.

Several impressive waves of alluviation have reached the Gulf Coast since the early Mesozoic, and each must reflect crustal unrest within the continental interior. During the late Mesozoic the volume of sediment transferred to the northern Gulf Coast was on the order of 160,000 cubic miles, nearly 100,000 of which arrived during the Lower Cretaceous. This wave was probably associated with events in the Appalachian and Ouachita regions. But what brought a Neogene wave which accounts for a section more than 20,000 feet thick and a volume of possibly 30,000 cubic miles? One's thirst for knowledge is not quenched by the explanation that the Gulf Coast simply subsided to accommodate the load. And are the fluctuations in volume of continental ice and stand of seas during the Quaternary sufficient to account for deposits some 6000 feet thick, with a volume on the order of 50,000 cubic miles? Each wave of alluviation must be the result of some pro-

nounced acceleration of denudation that occurred back on the continent.

The stratigrapher and sedimentologist are most concerned with the consequences of erosion. In deciphering the record, however, they should bear in mind the probability that the earth's landscapes, together with all erosional and depositional processes, have undergone at least four great changes which were related to the vegetational cover of the land, and hence to soil development (Russell, 1956, p. 454). The naked earth which existed prior to the middle Paleozoic certainly was without soils in the modern sense. Only incipient soils could have existed beneath the cover of primitive plants which existed before the late Mesozoic. A few modern soils appeared under Cretaceous forests, but it was not until grasses appeared in the early Tertiary that the remaining "great groups" of soils evolved.

Most significant, in the evolution of modern soils, are the changes which took place in the subsoil, because that part of the profile provides the most effective armor against weathering and erosional attack. Hill crests became rounded only as subsoils developed. An increasing complexity of clay minerals originated in the subsoils. Over much of today's earth the characteristic landforms result from morphogenic processes which are relatively recent in origin because they reflect the presence of subsoils. The Cenozoic may well be divided into an Eogene, during which modern landscapes started to develop, and Neogene, during which nonbiological factors such as increasing orogeny and volcanism, together with increasingly sharp climatic differentiation and zonation, which finally culminated in an ice age, completed the change to the earth's surface we know today. Never before has the earth been so well armored against processes of weathering and erosion. Effects of this armoring must exist in sedimentary deposits to a degree which makes it a bit hazardous to press too far any interpretation of the remote past on the basis of the present.

Alluvial Morphology

Concept of Recent

The Recent is the primary domain of the alluvial morphologist. To understand features such as alluvial valleys, deltaic coastal plains, and alluvial coasts, he needs three-dimensional information. Many of his techniques are those of the stratigrapher.

As a stage of the Quaternary, Recent is best defined as the time during which sea level made its last general rise, a usage dating from 1872 (T. M. Reade; C. Ricketts; A. Tylor). The oldest Recent marine sediments accumulated beyond the shores of seas at least 450 feet lower than at present, and lie well out from most coasts. Oxidized surface materials were progressively covered by a marine section while Recent seas were rising, and deposition advanced inland and upward. If the rise continues, not only today's Recent deposits but additional continental surfaces will become blanketed, possibly to the level of existing 200-foot contours (Ahlmann, 1953, p. 3). The record of the Quaternary suggests that this will not be the case, but the future is open to speculation.

To start the Recent with the initiation of the general rise of sea level is sound stratigraphic practice. Out on the continental slope and beyond, deposition has been uninterrupted, as has generally been the case throughout geologic time. A hiatus marking the interval between withdrawal and return of the sea is present across the shelf. Extensive gravel deposits commonly occur toward the base of the Recent, both along valleys and across deltaic facies of the shelf, providing a recognizable basal conglomerate. Local warping and tilting along coasts such as that of Louisiana establish an angular unconformity between the base of the Recent and whatever it overlies. These are fundamental arguments. Alternative possibilities, such as the time the sea reached comparative stillstand (Frye and Leonard, 1953), the beginning of the ice-free lake sequence in central North America, or the time the ice cleared the north German Coast, useful as they may be locally, are not fundamental in nature (Flint, 1947, p. 205–214). A similar line of reasoning relates the initiation of the Quaternary to the first appearance of continental ice (Russell, 1950).

With this concept of Recent in mind, we may now turn to some of the problems of alluvial morphology. The scope of the discussion is restricted to problems which are essentially geological and according to limits of my own familiarity with the field.

Flood-plain Deposits

Seaward portions of most main river valleys establish the universality of Recent drowning. Where there is an abundance of sediment the drowning may be alluvial, or deltas may pro-

trude short distances beyond adjacent coast lines. If adequate subsurface information exists an irregular, pre-Recent topography may be identified below the alluvial fill. The thickness of the alluvium increases notably below deltas.

Grade sizes of sediments vary according to local circumstances but ordinarily are finer in upper parts of the alluvium. The delta of the Rhone, the Camargue, is surfaced by rather coarse sand, but increasing amounts of gravel occur at depth (Russell, 1942, p. 234–236). The Lower Mississippi transports only small amounts of fine sand to its delta, but practically all samples from the lower Recent in the vicinity contain some gravel. An alluvial sequence in which coarse materials underlie fine is practically universal, not only in Recent alluvium but also in the Pleistocene terraces which occur along larger valleys. In many cases there is a fairly sharp break between the coarse and fine parts of the section.

Complications such as conspicuous lenses of coarse sand or gravel in otherwise fine-grained materials are common in river alluvium. Many appear to be related to the scouring of pools, during floods, to depths where the coarse, basal materials are located. Erosion detaches grains or pebbles which are transported turbulently upward and deposited on banks or bars located well above low-stage levels of the river, where they become incorporated in finer deposits.

While there is commonly a general decrease in grade size of sediments downstream, there are likely to be more pronounced contrasts within short distances across valleys. Coarse alluvium is concentrated along the natural levees of meander belts, particularly in point bars, whereas very fine silt, clay, or organic ooze may be accumulating in near-by backswamp basins.

The meandering process mixes sediments to such a degree that distinctive suites of minerals which may characterize tributary streams disappear through dilution in short distances on the main flood plain (R. D. Russell, 1937, p. 1333). Swinging and sweeping of meanders may account for some of the mixing, but the dominant process involved depends on the fact that sediments ranging upward in size from fine silt rarely travel more than the distance from a scour pool to the next bar downstream, where they lodge quite permanently (Matthes, 1941).

The freedom of a meandering river to wander is ordinarily exaggerated. The enlargement of a loop is usually limited by the development of a neck cut-off. Cut-off lake distribution normally suggests widths of individual meander belts, as along the Lower Mississippi in the vicinity of Greenville (Fig. 1*A*).

The lake in an abandoned crescent gradually fills with fine sediment and becomes a clay plug (Fisk, 1947), which is nearly unerodible. Each new cut-off normally returns a river to a central position in its meander belt, and each clay plug tends to hold it there (Fig. 1*B*). New meander belts result from major diversions, and these are rare on many rivers. During the last 6000 years not more than half a dozen have originated in the Lower Mississippi Valley, and the typical width of each has been less than one quarter that of the flood plain (Fig. 1*C*). Some of the large, inter-meander-belt basins bear no surface trace of main channel occupance.

Channel Types

General Discussion—The most typical meandering rivers vary in pattern and cross section as parts of their channels, in turn, are fixed, anastomotic, or meandering. These principal channel types are related mainly to differences in hydraulic gradient and contrasts in bed materials.

Fixed channels are characteristic of eroding streams, particularly in rugged territory. Bed materials are eroded and transported out of the region. Pools or shoals develop according to the erodibility of channel beds or sides. Significant shifts in channel position take place slowly, as in the case of lateral downdip migrations which produce asymmetrical valleys.

Streams in alluvial valleys also develop fixed channels by eroding their beds into consolidated bedrock or tight clay. The cause of channel fixing may not be apparent at the surface, but the effect may be as definite as if the channel had been superimposed across a rocky hill. Much of the channel of the Lower Mississippi between Baton Rouge and New Orleans is fixed in compact clay and silt of a Pleistocene terrace formation (Fig. 1*D*). River distances have not changed a quarter of a mile during the last 2 centuries, nor are there any traces of abandoned meander loops on the adjacent flood plain. The meanderlike kinks along the river are dominantly controlled by fractures.

The Lower Mississippi course below New Orleans is complicated by the Lake Borgne fault zone (Fig. 1*E*), below which the channel becomes fixed in tight clay which was deposited ahead of a much older delta, that of the Teche-Mississippi (Russell, 1940). This fixed channel is comparatively straight. At various places

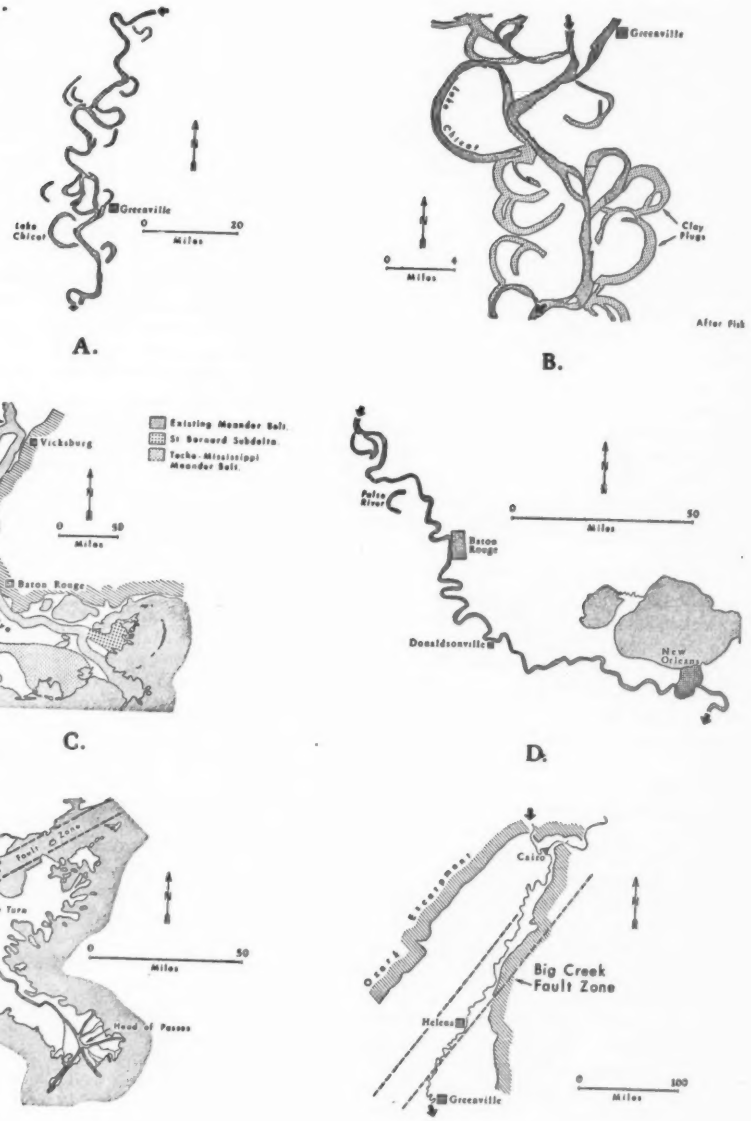


FIGURE 1.—LOWER MISSISSIPPI STREAM PATTERNS

A. Meander belt with cut-off lakes, Near Greenville, Miss.
B. Clay plugs and cut-off lakes along meander belt.
C. Two latest meander belts in southern Louisiana.
D. Fixed channel below Baton Rouge.
E. Straight channel below New Orleans.
F. Fixed channel along Big Creek Fault Zone.

upstream the Lower Mississippi is fixed along fault zones, most striking of which is the Big Creek fault zone, which strikes northeastward in the vicinity of Helena, Arkansas (Fig. 1*F*). In local instances the channel has eroded its bed into compact or consolidated materials—particularly into clay plugs or Tertiary bedrock at slight depth—which fix it firmly.

The most characteristic patterns of meandering rivers are anastomotic and meandering. The braided, anastomotic channel superficially appears to be the more complex, with its numerous branches, lenticular islands, and bars. The meandering channel is comparatively simple in plan, with fairly regular sinuosity and modest number of bifurcations. Variations from one pattern to the other ordinarily depend on changes in hydraulic gradient or the character of bed materials. Steeper gradients and less consolidated, more erodible, and coarser alluvium favor anastomosing. But similar patterns occur in deltas, where gradient decreases abruptly. The fundamental cause in all cases is surplus load.

Stream patterns originate during higher stages of flow. With rising stage, velocity and turbulence increase, pools are scoured if bed materials permit, and erosion is most active. It is erroneous to believe, however, that times of rising stage are characterized by erosion alone, or that deposition is restricted to falling or low stage. A better generalization would recognize increasing discharge as accelerating both erosion and deposition. The distance is ordinarily short between a site of active erosion and a place where deposition is vigorous. It is the distance from a scour pool to some near-by part of the channel where turbulence falls off rapidly, or where water is actually slack.

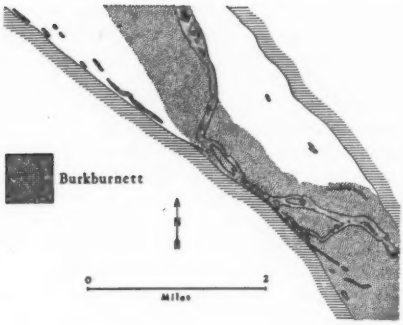
An anastomotic river, such as long distances of Red River between Texas and Oklahoma (Fig. 2*A*), erodes its bed vigorously at time of flood. At the same time it is building bars and creating a highly irregular bed topography. As a flood subsides, turbulence and erosion, turbidity and deposition all diminish in intensity. When low stage is reached there is little or no activity along the bed. When most anastomotic in appearance, at lowest stage, a river wanders through the maze of irregularities its bed has inherited from the activities of the preceding high stage.

In readily erodible alluvium channels tend to widen and shoal and develop anastomotic patterns (Leighly, 1932; 1934). In highly resistant materials they tend to be fixed. It is an intermediate condition which permits meandering. For typical Lower Mississippi River hydraulic gradients meandering is favored by beds of medium or fine silt, or silt admixed with moderate amounts of clay. Essential to the meandering process is a bed into which pools may be scoured, associated with banks which remain fairly firm and steep at high stage.

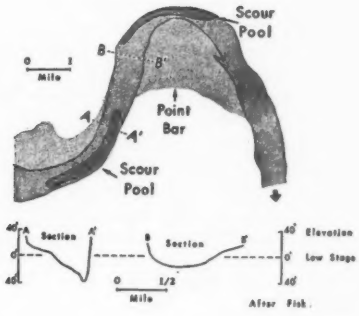
The scour pools beneath concave banks of meandering channels deepen with rising stage (Fig. 2*B*). Detached sediments are promptly deposited on the next bar down-stream (Matthes, 1941). The accretion increases the volume of the bar, which grows out into the channel, producing a tendency toward diminishing channel cross section, and thus increasing hydraulic gradient, velocity, and turbulence. To accommodate its discharge the river deepens the scour pool at the toe of the bar. When deepening of the pool weakens sufficiently the foundational support of the concave bank opposite the bar, bank materials subside, slump, or slide off into the channel, to be carried downstream to the next bar on that side of the channel (Fig. 2*C*). It is in this way that a "cut bank" recedes (Fisk, 1944; 1952a; Russell, 1954a, p. 16) and the bend of a meander loop is accentuated. In cases where a cut bank is protected by a revetment, the scour pool normally deepens to such a degree that the revetment eventually slides into the pool. If meandering is to be inhibited, it would probably be more effective to pave scour pools than to revet concave banks. Revetments appear to be designed with the idea in mind that lateral corrasion causes recession of the bank. This concept, while widely taught and plausible when speculating about meandering on the basis of map evidence alone, is unsound. One will never see debris thrown against cut banks during floods on large rivers. The surface water there is commonly slack, or actually flowing upstream. Erosional atack is concentrated in the scour pools below.

Local steepening of hydraulic gradient promotes local bed erosion. The effect is similar to that of providing a bed of more erodible alluvium and thus leads to channel widening and development of anastomotic pattern. Along the Lower Mississippi River abnormally long reaches, towheads, or chutes ordinarily mean that the gradient has steepened locally, or that bed materials are coarser (commonly sandy), more heterogeneous in size, or lack clay content. These conditions are more prevalent above than below Helena.

The lower Rhone alternates between meandering and anastomosing in relation to cones of



A.



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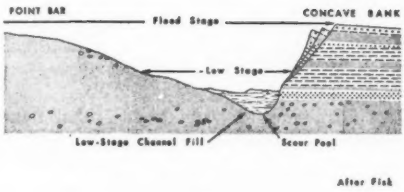

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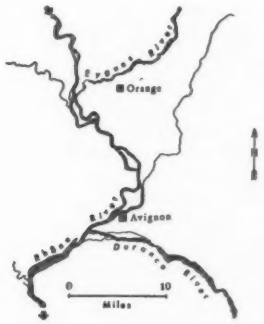

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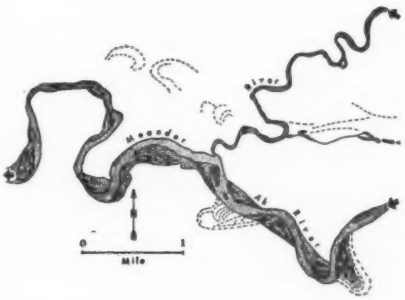

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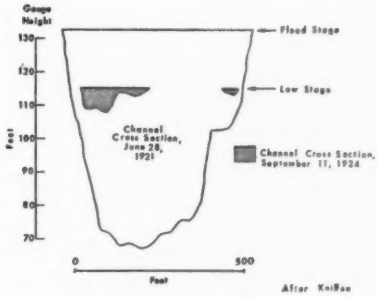

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FIGURE 2.—ANASTOMOTIC AND MEANDER PATTERNS
A. Anastomotic channel of Red River, Burkburnett, Texas.
B. Essential parts of a meander.
C. Section of channel at "cut" or "concave" bank.
D. Meandering and anastomosing on the lower Rhone.
E. Meandering and anastomosing on the Great Meander.
F. Extreme channel sections, Colorado River at Yuma, Arizona.

dejection of its tributary streams. On gentle gradients upstream from the cones it meanders with some success, but on downvalley slopes of the cones the gradient steepens, to increase turbulence and scour, so that the river shoals and flows in divided channels (Fig. 2*D*). An interesting example of the effectiveness of these processes occurred along the Lower Mississippi following the Engineers' cut-off program of the 1930's. In the cut-offs, hydraulic gradients steepened to a degree where shoaling became so pronounced that dredging was required to maintain channels of navigable depth, both in and below cut-off's.

Sediment contributed by tributaries also affects stream patterns. No finer example exists than along the Great Meander River of western Anatolia, the type locality for meandering streams. For much of its upper course the Meander follows a graben along which overwhelmingly more active alluvial cones from the north fix the channel against the south valley wall. Near Nazilli the river escapes from fixed-channel confinement and reaches the head of its alluvial plain. Here the alluvium is notably less gravelly, the gradient flattens, and the Meander starts its characteristic sinuosity. But farther downstream such tributaries as the Ak and Çine, which resemble Arizona washes in most essentials, bring overwhelming loads of coarse detritus that abruptly change the pattern of the master stream into an anastomotic river for various distances below their junctions (Fig. 2*E*; Russell, 1954b, p. 366–374).

During the normal seasonal cycle between low and flood stage many rivers tend to change from meandering to anastomotic streams. Floods increase the hydraulic gradient and straighten out threads of maximum velocity. Captains of river boats are keenly aware of this. For the sake of fuel economy boats headed upstream are routed as close as possible to bars, but during floods they hug concave banks. As rivers straighten during floods they commonly shunt across bars and are likely to develop chute cut-offs. Channel bifurcations on bars are distinctly anastomotic in pattern. Most meandering rivers, could they be maintained at flood stage for some years, would become relatively straight, broad, shallow, anastomotic streams. The converse is not true. If kept at low stage indefinitely, a river would continue to flow along whatever channel it happened to inherit, without ability to scour pools, build bars, or change its channel section appreciably.

Channel sections are highly sensitive to demands imposed by discharge. A measured section across the Colorado River at Yuma, Arizona, presents an extreme case (Fig. 2*F*; Kniffen, 1932, p. 164). As normal routine, section enlarges with increasing discharge and diminishes as stage drops. Erosion or sedimentation occurs according to changes in discharge, but perfect equilibrium between discharge and section is not likely to occur along a river because there is a lag in morphological accomplishment. About the time that equilibrium with low-stage channel demand is approached, for example, sediments which have recently been deposited start eroding away if the stage starts to rise. One aspect of this whole question is the volume of the "reservoirs" of silt that commonly lie along channel banks, only to disappear during the next flood.

In the case of a river which has been diverted into a new course, the initial route is determined by the topography of the basins it crosses. If the alluvium is readily erodible, the new channel develops an appropriate section within a comparatively short time. Early in the process an anastomotic phase is initiated as a rule, but if bed conditions are suitable meandering will develop later.

An abandoned course, on the other hand, undergoes a history of channel deterioration which depends on the supply of sediment it receives. If the diversion is abrupt and a new natural levee is built rapidly across the head of the abandoned course, the detached channel may remain comparatively unaltered for centuries. More commonly there is some flow down the abandoned course for many years, at least during times of flood. This flow brings sediment which effectively diminishes channel section in accordance with the requirements of the lessened discharge. Cane River, Louisiana, is a good example (Fig. 3*A*). Prior to 1836, it was the main channel of Red River below Natchitoches, but a diversion that year resulted in its abandonment. As a result of continued flow along the abandoned channel sufficient sediment was furnished for channel deterioration which, in less than a century, reduced the section so that it could not handle 10 per cent of the normal discharge of Red River.

The Atchafalaya River of southern Louisiana illustrates the morphological effect of constantly increasing discharge. In 1860 removal of a raft of vegetational debris and various improvements were initiated for opening the Atchafalaya to navigation and making it a more effective distributary of the Lower Mississippi.

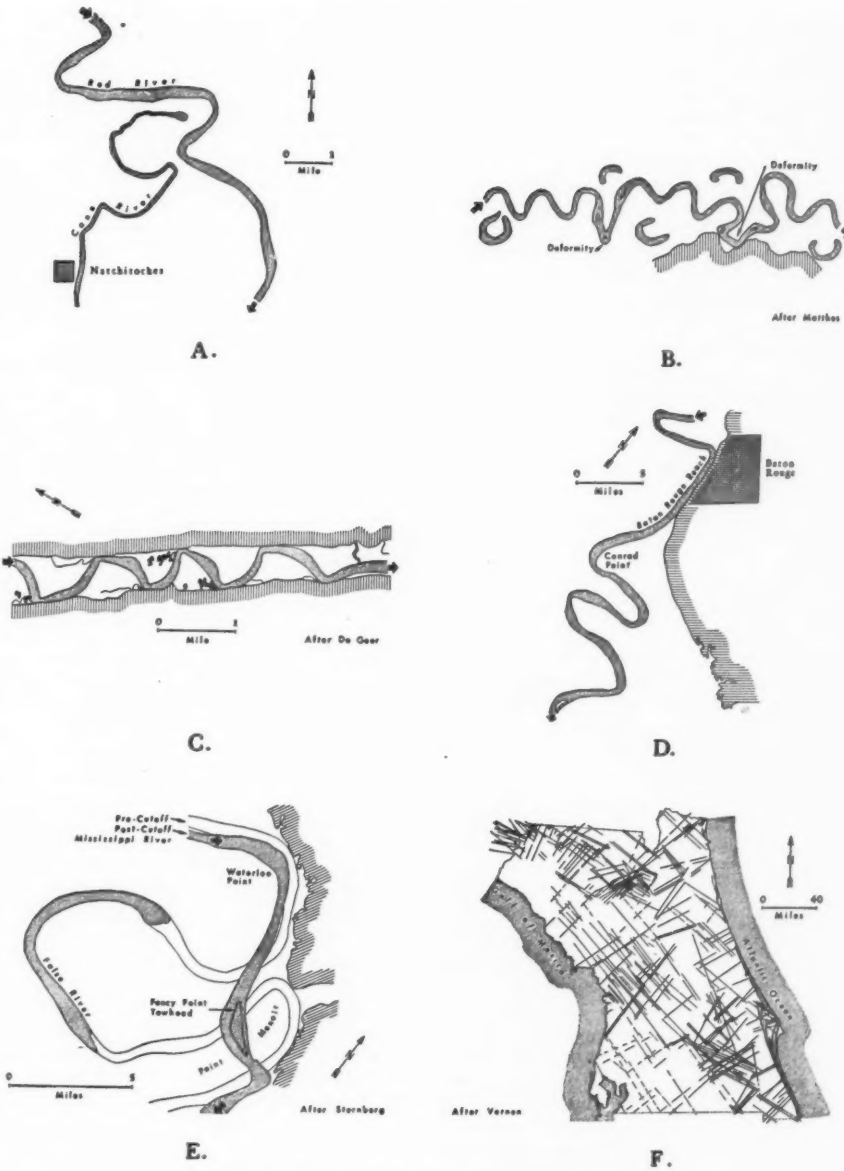


FIGURE 3.—FACTORS AFFECTING CHANNEL PATTERNS
A. Existing and abandoned channel of Red River, La.
B. Idealized meander belt, with deformities.
C. Klarälven River, Sweden, between narrow valley walls.
D. Baton Rouge Reach of Lower Mississippi, at valley wall.
E. False River, La., and points of adjacent loops
F. Fault patterns in northern Florida.

Since then the discharge has increased, progressively, during every decade, so that the channel now carries more than a quarter of the flood discharge of the main river. Increasing discharge has demanded channel enlargement which has never been fully achieved. In firm clay the river is both narrow and shoal. Where the bed is locally more erodible the channel is both wide and deep. This is a type of channel immaturity typical of new river courses, such as initial channels below points of diversion. The upper Atchafalaya has attained a pattern of incipient meandering, but the lower river is anastomotic and has not yet alluviated its basin to a point where its pattern has overcome control by initial topographic irregularities.

Deformities—Even where all factors favorable to meandering exist, most rivers exhibit interruptions in pattern by abnormal reaches, distorted loops, or unexpected changes in course direction. Gerard H. Matthes (1941), master student of river behavior, applied the term, deformities, to these pattern interruptions (Fig. 3*B*).

A fundamental type of deformity occurs along part of the River Klarälven in southern Sweden (DeGeer, 1906, p. 388–404; Sundborg, 1956, p. 244–292). Meander loops appropriate in size to the discharge are prevented from developing by narrow valley walls (Fig. 3*C*). The start of each loop initiates a radius of curvature which might more than double the width of the valley floor, a pattern impossible of realization. On impingement against a valley wall the river is directed abruptly into a down-valley reach, after which it attempts development of its next loop, with the same result.

Similar deformities occur along some rivers of the Allegheny Plateau and elsewhere. Curiously enough, they have been characterized in literature as "underfit," in the erroneous belief that the valley flats approximate the widths of rivers which were supposed to have created them erosionally. Actually the valleys are alluvially filled, and their rivers more appropriately might be thought of as "overfit," because their flood plains are too narrow to permit complete expression of their meandering potentialities.

The Baton Rouge Reach of the Lower Mississippi is a related case of channel deformity, resulting from impingement against a valley wall (Fig. 3*D*). The river strikes directly a Pleistocene terrace, with silty clay at bed level which can be eroded only with difficulty, so that meandering is inhibited, and the channel is deflected into the reach. About 7 miles downstream there are silts at bed level which were deposited along an earlier, modern course, where the river scours, gains in turbidity, and starts meandering, around Conrad Point. Not far downstream, however, meandering yields to fixed channels, in Pleistocene clay.

The most subtle cases of deformity result from channel contacts with clay plugs or bedrock which lie blanketed under alluvium. These interruptions in ability to scour or enlarge the channel section are as effective as valley walls in terminating loops and introducing abrupt changes in patterns, which normally take the forms of reaches or acute bends.

The overdevelopment of a particular meander loop is a type of deformity illustrated by False River, a short distance upstream from Baton Rouge. This is not only the last cut-off downstream on the Mississippi, but it is also the longest (Fig. 3*E*). The explanation of its great length, which originally was well over 20 miles, depends on the fact that adjacent meanders both upstream and downstream, attained fixed positions against the eastern valley wall, where they were held for an extreme length of time, during which False River developed its abnormal length (Sternberg, 1956).

Many deformities are structural in origin. In the most common case rivers are trapped along faults or follow trends of grabens.

Tectonic Influences

Some years ago, when Dr. Harold N. Fisk and the writer began noting on aerial photographs straight or gently curvilinear lines extending across varying types of terrain and vegetational cover, we were so appalled by their numbers that we were inclined to regard them as an expression of jointing. But the test of boring closely spaced holes across their strikes invariably revealed offsets which increased progressively with depth, and it became clear that the lines on the photographs were actually fault traces. We also found cases where the traces continued as faults in bedrock beyond valley walls. Lower Mississippi Valley fault patterns resemble closely similar features in Florida (Fig. 3*F*; Vernon, 1951, p. 47–52) and in other parts of the world.

Among interesting examples of Lower Mississippi Valley faulting is the fracture which separates Reelfoot Lake Basin from Tiptonville Dome. This fault undoubtedly was responsible for the New Madrid earthquakes of 1811–1812. The surface throw amounts to about 20 feet,

but 10 borings within a mile, of which 5 were concentrated within 1000 feet, were sufficient to demonstrate drag effects near the fault surface and a throw of nearly 40 feet at the top of the Tertiary (Eocene). The section has been published by Krinitzsky (1950).

Faulting determines both gross and localized positions of Lower Mississippi River channels. Though the river meanders at many places above Helena, its general trend is along the Big Creek fault zone (Fig. 1*F*). This region is seismic, as evidenced by numerous small earthquakes during recent years. Every major diversion of the river has occurred on a fault, as have most of the small crevasse channels which lead across natural levees into adjacent basins.

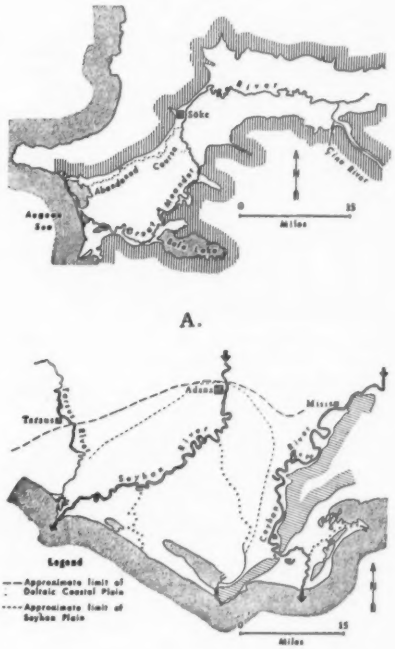
A curious example of faulting at Vacherie, about 25 miles west of New Orleans, resulted in a maximum throw of 14 inches during activity which lasted for 10 days in April 1943. No earthquake was reported, but a remarkable pattern of en echelon fractures developed along a zone about 100 feet wide and 2.5 miles in length. It was apparent on aerial photographs that the site had been one of persistent flood crevassing before it was protected by an artificial levee, and it seemed probable that the displacement was caused by differential compaction of sediments along the crevasse channel. Boring demonstrated, however, a throw which increased progressively to 30 feet within a depth of 900 feet along this particular fault.

The rotation of downthrown blocks along normal faults commonly tilts alluvial surfaces toward valley walls, so that channel positions appear anomalous with respect to growing alluvial cones or colluvial aprons. This relationship was noted in Carson Valley, Nevada, by Lawson (1912), who observed,

> "The Carson River flows northerly at the base of the mountain and is distributed through an extensive marsh, the waters of which are in places so close to the rocky slope that one may stand upon the latter and throw a stone into them with ease."

Nowhere is the effect of rotation better demonstrated than along the Great Meander River. An old, well-preserved meander belt hugs the northern wall of a southwestward-trending graben for a distance of about 25 miles (Fig. 4*A*). A recent diversion directed the river southward near Söke, to the south graben wall, which the Meander now follows for more than 20 miles (Russell, 1954b, p. 370–374).

Another Anatolian example of tectonic control of rivers is the striking contrast between the Seyhan and Ceyhan rivers as they flow across their common deltaic plain toward the northeastern corner of the Mediterranean Sea (Fig. 4*B*). These rivers have about the same discharge, flow across similar alluvium, and



A.

B.

FIGURE 4.—TECTONIC CONTROL OF RIVER COURSES, ANATOLIA

A. Ancient and existing courses of the Great Meander. B. Contrast between Seyhan and Ceyhan deltaic plains.

exhibit similar flood characteristics, but their patterns have little in common. The Seyhan wanders freely over most of the deltaic plain and has left an impressive record of abandoned channels and meander scars, whereas the Ceyhan has been trapped near the base of a range of hills, where it flows along a comparatively straight meander belt, on a rotated, downthrown fault block.

A somewhat similar relationship exists north of the Bay of Bengal, in the common deltaic plain of the Ganges-Brahmaputra (Fig. 5). The Ganges has had relative freedom in establishing outlets on the western side of the plain. Trunk channels were initiated and subsequently

abandoned as a result of upstream diversions. To the east, the Brahmaputra courses lie in an area of tilted blocks, where active faulting controls channel positions. Deltaic sedimentation is concentrated in comparatively restricted

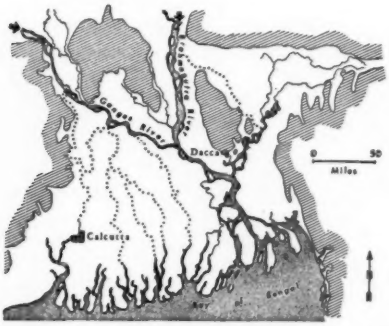


FIGURE 5.—CHANNELS OF THE GANGES AND BRAHMAPUTRA, BENGAL

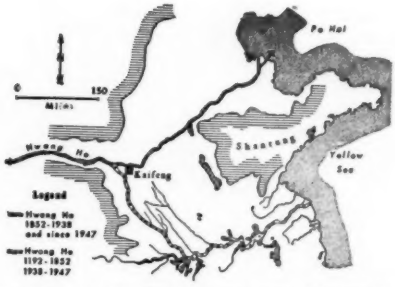


FIGURE 6.—DIVERSIONS OF THE HWANG-HO LEADING TO DELTAS ON OPPOSITE SIDES OF SHANTUNG PENINSULA

areas along subsiding troughs. Structural activity has recently caused capture of the Ganges by the Brahmaptura, leaving the western part of the deltaic plain abandoned, so that former mouths of the Ganges remain only as tidal estuaries (Morgan and McIntire, 1957).

Deltas

The site of any particular delta depends on the topography upvalley. Any new river course is determined by the arrangement of basins below a point of diversion. The fact that the deltas of the Hwang-Ho have grown alternately on either side of Shantung Peninsula stems from diversions which originate about 250 miles inland (Fig. 6). A succession of modern deltas of the Lower Mississippi involves events leading to the creation and abandonment of meander belts by diversions occurring several hundred miles from the coast, in eastern and northern Louisiana. It is more than 250 miles upstream to the point where the Atchafalaya River now threatens the Lower Mississippi (Fisk, 1952b), yet this diversion, if accomplished, will be located farther downstream than the head of any other major meander belt.

Meander belts that originate well inland are entirely different from the relatively short distributaries of deltas, such as the "passes" of today's bird-foot delta or fan-shaped courses of still older Lower Mississippi River deltas. These coastal outlets exhibit patterns which originate as submarine features.

As the mouth of an alluviating river advances seaward, the channel normally widens, and its natural levees tend to flare and diverge, both above and below sea level (Fig. 7). Natural levees grow outward across shallow bars which lie beyond terminal tips of dry land. The whole system moves seaward as the delta's land framework is extended. Each channel mouth widens and shoals as the river bed slopes upward, toward the bar. Active deposition on the bar is controlled by irregular patterns of turbulence and abilities of currents to transport suspended load and bed load. Shoals and secondary bars develop where deposition is most rapid. One of the preferred sites of shoal location is mid-channel in flaring outlets.

Submarine natural levees grow seaward from the more stable shoals and bars at the mouths of distributaries; their patterns resemble the clavicles of a wishbone. These alluvial ridges extend outward under water, and the channels they confine commonly develop submerged mid-channel shoals. With increasing distance from the shore a pattern of underwater channel bifurcations develops in complexity with something like geometrical progression. This evolution of distributary patterns is well illustrated in the area east of Cubits Gap, Lower Mississippi River Delta, where the land is now advancing rapidly into the Gulf of Mexico. Detail sketched from aerial photographs is presented in Figure 7*A*, but visibility limitations obscure the more complex patterns at depth. Vicissitudes of the environment result in modification and destruction of many shoals, channels, and natural levees, but those that survive eventually determine a delta's channel, land, marsh, and bay patterns (Welder, 1955).

The process of underwater channel bifurca-

tion is illustrated beautifully at the mouth of the Sakarya River as it enters the Black Sea, in northwestern Anatolia. An abundance of sand is transported by the river, even at low stage. New channels, submarine natural levees, and

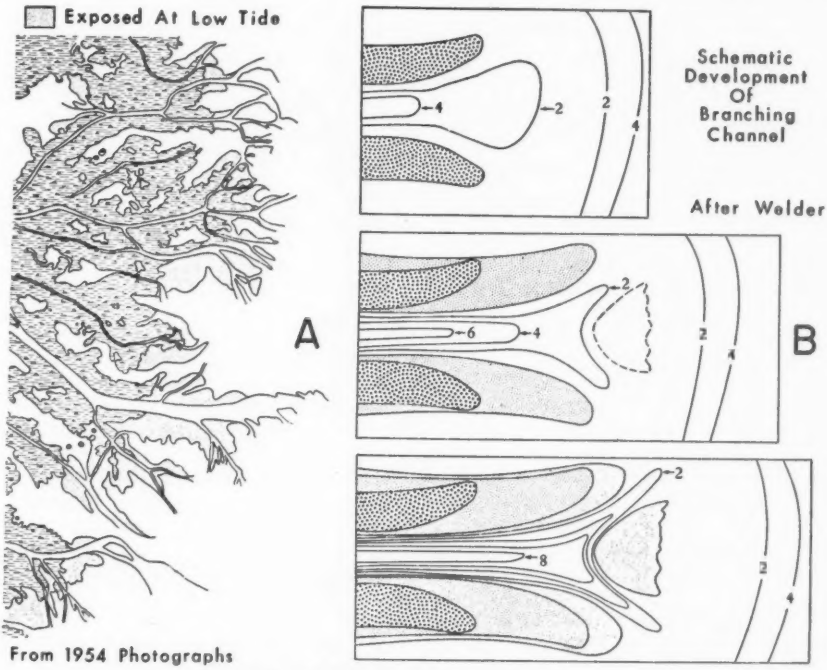


FIGURE 7.—ORIGIN OF DELTA STREAM PATTERNS
A. Land, tidal-flat, and channels traced from photographs, Cubits Gap area, La.
B. Schematic development of shoals and channels.

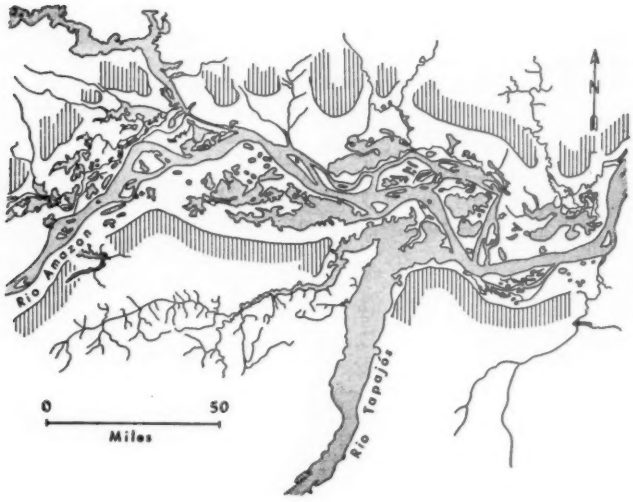
shoals develop so rapidly that the services of a pilot are required throughout the day, in order that boats which set out in the morning may be guided back into the river mouth later in the day (Russell, 1954b, p. 364–366).

The pattern of submarine natural levees is essentially anastomotic. Many channels which separate at the inland tips of wishbone-shaped bars or islands later join or reach neighboring channels. For this reason many of the marshy islands of deltas are lenticular and sharply pointed at each end. The closing of less well situated channels at bifurcations as a result of bar building in slack water leaves many "blind passes" or channels beheaded by deposition.

From somewhere above Manaus to the vicinity of the Xingu the Amazon exhibits delta rather than flood-plain channel patterns (Fig. 8*A*). A few thousand years ago the valley of the main river and its tributaries was a large, complicated estuary, with arms extending across much of northern Brazil. Alluviation along the main channel has progressed so rapidly that many of the tributary arms of the estuary have been dammed off and converted to elongate, wedgelike lakes, which are widest at their lower ends. The Tapajós and Xingu are excellent examples.

The lenticular islands of the Amazon deltaic plain are in many cases little more than willow-fringed shells which enclose lakes or grassy flats that become inundated during annual floods. The maze of channels surrounding these islands exhibits patterns which were created under water, during the advance of the delta into the original deep estuary. Most main channels are straight or gently curvilinear. They duplicate patterns of Lower Mississippi River delta distributaries below Cubits Gap.

In sharp contrast to the patterns developed by the main channels of the Amazon deltaic plain are those of innumerable minor steams that lead across natural levees into the basins



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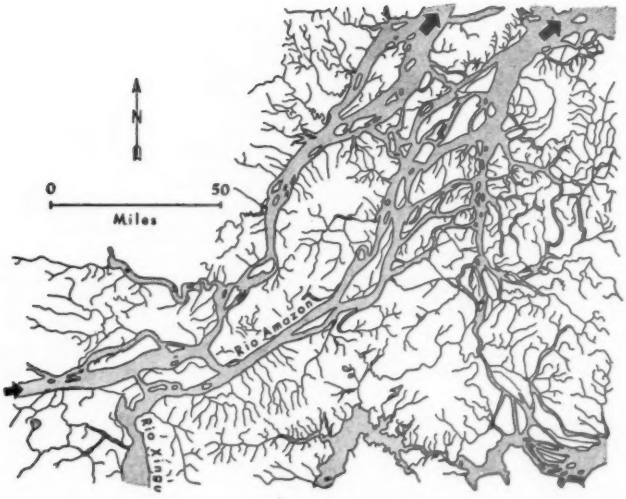

B.

FIGURE 8.—CHANNEL PATTERNS OF THE AMAZON
A. Delta pattern of the filled estuary.
B. Archipelago pattern toward the river mouth.

within the lenticular islands. These streams meander tightly, with bends and loops of sizes appropriate to their discharges. Unlike the through channels, the smaller streams are creating their courses across alluvial land where they can scour pools, build bars, and thus develop true meandering. Similar processes result in the meanders of tidal streams in the Lower Mississippi and other deltas, and in coastal marshes generally.

Below the Xingu the Amazon leaves its deltaic plain and flows across a much older archipelago (Fig. 8*B*). The islands here are high, flat, and either extensively covered by forest or cleared for agriculture. Rims of dead and leaning trees fringe shores which are being eroded actively. On some islands two or more terrace surfaces are evident. This river-mouth region is being uplifted and in this regard resembles a highland with elevations in excess of 400 feet which parallels the Gulf Coast along the Louisiana-Mississippi boundary, immediately north of the site of most active deposition of latest Mississippi River deltas (Russell, 1940, p. 1218–1219). Both areas suggest uplift resulting from isostatic adjustment. Both display uplifted, depositional Pleistocene terraces. The upwarping in the Lower Mississippi Valley involves terraces which slope abnormally steeply southward, toward the delta, and gently northward toward Jackson, Mississippi (Fisk, 1939). Has the loading of the shelf off the mouth of the Amazon elevated the river-mouth region in a similar manner?

The outlines of recent Amazon history appears to be related to Quaternary changes of sea level. Valleys were cut during low stands of the sea and were filled during times of waning glaciation, as seas rose. Dissection of the coastal belt, which produced the main valleys between islands of the archipelago, probably resulted from erosional activities during one or more low stands of Pleistocene seas. The surface patterns upstream, in the estuarine delta, on the other hand, are so fresh and are now changing so rapidly that they must be related to the stillstand of high sea level during the past 5000 years or so. Also of Recent origin is a submarine delta which is being built on the shelf, seaward from the archipelago.

The bird-foot delta of the Mississippi is less than 500 years old (Fisk, McFarlan, Kolb, and Wilbert, 1954, p. 97). Its pattern depends on the fact that the river is confined to a single channel for about 70 miles below the Lake Borgne fault zone (Fig. 1*E*) by compact clay. Through this clay the channel is deep, flat-floored, and comparatively straight. Beyond the clay the river encountered fairly deep water where it deposited coarser alluvium and formed its distributaries in bar sand (Fisk, 1955), initially as submarine features.

Older deltas of the modern Mississippi generally resembled the delta of the Nile in shape (Fig. 9). The St. Bernard Delta, east of New Orleans, contains wood samples which have been dated back to a maximum of about 2000 years by C^{14} analyses. This delta developed eastward from the area dominated by Teche-Mississippi Delta clays which were deposited some 4000 years ago. Complicated channel patterns southwest of New Orleans are related in part to pre-Teche-Mississippi courses and in part to more recent distributaries which crossed coarser facies of the Teche-Mississippi Delta, where they were able to spread out fanlike.

Deltaic Coastal Plains

The alluvial advance of coasts which are well supplied with river sediment forms lowlands which Barton (1930) aptly termed, deltaic coastal plains. These flats may be associated with the activities of a single delta, with a succession of deltas of the same river, or may be the compound product of many rivers, as is the case along the northern and western shores of the Gulf of Mexico.

The highest and firmest land of a deltaic coastal plain consists of natural levees which flank the channels responsible for the lowland. Basins which exhibit various degrees of filling lie within the natural levee framework. Some contain lakes, others grassy marshes or tree-covered swamps. Secondary and minor channels lead into the basins from levee crests and also to adjacent bays or the sea. Each basin ordinarily develops a drainage system of its own (Russell, 1939, p. 1221–1227). Anastomotic patterns of the Amazon type, here and there, reflect a history of lake or bay extinction, such as has occurred in the lakes of the Atchafalaya Basin or Garden Island Bay, east of South Pass, in the bird-foot delta, during the past 60 years (Russell, 1936; 1938).

The rapidity with which a deltaic coastal plain may develop is well demonstrated in western Anatolia, where the Little Meander has advanced the shore line more than 4 miles in less than 20 centuries, and the Great Meander least 10 miles in 25 centuries (Russell, 1954b, p. 374–376). The Great Meander filled the ancient Latmian Gulf so rapidly that one arm

of its former estuary was alluvially dammed to become Bafa Lake (Fig. 3*A*), which is reported to be 30 fathoms deep and appears to retain some of its original salinity.

An interesting and fairly common modification of deltaic coastal plains develops in cases where the supply of river sediment is so variable that coastal advance, which occurs when the supply is abundant, alternates with coastal retreat, when the supply is deficient and erosion gains dominance. As a rule these alternations in sediment supply result from shifts in positions of river outlets along the coast. During times of coastal retreat beaches are formed and pushed inland at the expense of coastal marsh or other alluvium. Coastal advance is accomplished by the growth of mudflats on the outer side of the latest beach, which in an inactive condition becomes a chenier (Russell and Howe, 1935).

The coast of western Louisiana has been retreating for several centuries, in places at a rate of 600 feet per century. A continuous beach was originated, consisting of sand, silt, shell, and coarse debris, which differs in composition and appearance from the underlying, overridden marsh deposits. In 1947, however, as a result of near completion of basin filling by the Atchafalaya River, sediment began to reach western Louisiana in sufficient quantities to initiate the development of extensive mudflats (Fig. 10). Marsh vegetation soon gained foothold on the original flats, so that the beach of 1947, where protected by mudflats, has halted in its retreat and now lies several hundred yards from the coast in many places. The mudflats have been growing at a vigorous rate westward (Morgan, VanLopik, and Nichols, 1953; Russell, 1953). Though some of the mud was washed back into the coastal marsh during the severe hurricane of June 1957, the mudflat survived both a hurricane wave at least 11 feet high and subsequent outpouring of excessive amounts of water from the marshes. There is

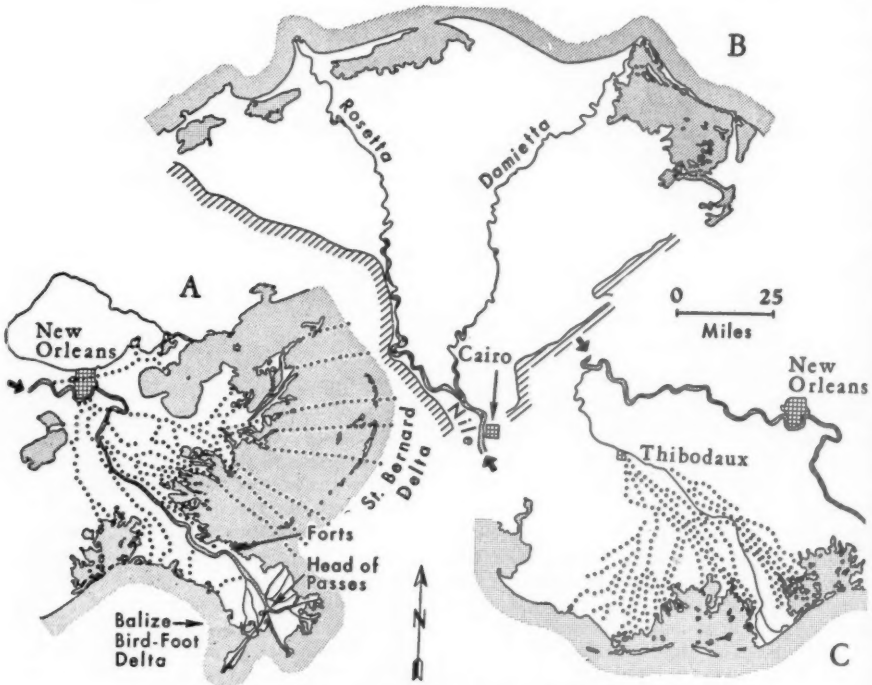

FIGURE 9.—DELTAS OF THE NILE AND LOWER MISSISSIPPI
A. Balize, bird-foot, and older St. Bernard deltas.
B. Nile Delta, on same scale.
C. Older delta complex, southwest of New Orleans.

little doubt but that the beach of 1947 close to the Atchafalaya outlet, and its continuation westward as the mudflat advances in that direction, is in process of becoming a chenier.

Price (1954) referred to coastal plains which are characterized by cheniers as chenier plains. Chenier plains are well developed in western Louisiana, Dutch Guiana, and elsewhere. Both the Rhone and Po deltas exhibit cheniers.

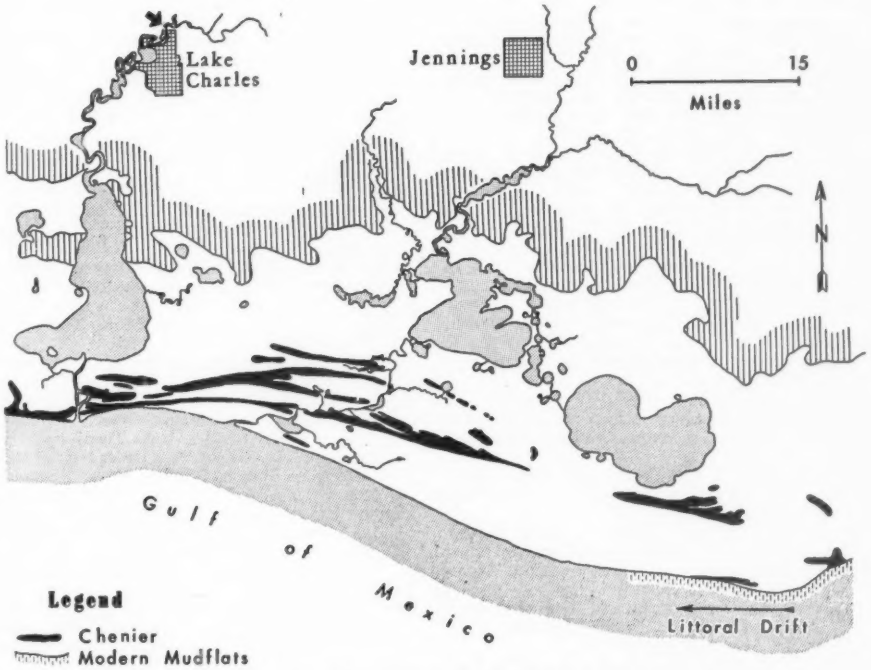

FIGURE 10.—CHENIERS OF SOUTHWESTERN LOUISIANA

CONCLUSION

The content of this address has been circumscribed both by time limitations and by emphasis of parts of geomorphology with which I am personally engaged. Nine years ago, in a presidential address to the Association of American Geographers, I attempted to outline ways in which geomorphology might better serve geographers (1949). This evening I have concentrated on topics which are closely related to stratigraphy and sedimentology. The coverage has been incomplete, even within the field of alluvial morphology, where subjects such as glaciated plains, alluvial aprons, and eolian surfaces have been omitted. Only incidental mention of Pleistocene surfaces has been included, and practically nothing about colluvial morphology. It is my hope, however, that this incomplete and brief statement may stimulate interest in the field and still further promote the interests of the Society's Geomorphology Group, which is now demonstrating a vitality suggesting notable progress toward the comprehensive chapter envisioned by Gilbert.

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CORONITES FROM INDIA AND THEIR BEARING ON THE ORIGIN OF CORONAS

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ABSTRACT

Detailed petrographic and mineralogic studies of some coronites from India confirm observations of previous workers on similar rocks. The first rhombic pyroxene corona is formed by replacement of olivine, and the second corona of amphibole, amphibole-spinel, and/or garnet develops by replacement of plagioclase. Certain features such as the clouding of plagioclase, pyroxenes, and olivines and the recrystallization of primary minerals (*e.g.*, pyroxene) and the development of garnet are commonly noted in coronites. Their significance as metamorphic phenomena and their relationship with corona formation is stressed. From a crystallochemical examination of the reacting minerals and the resulting corona minerals, the writer concludes that the first corona of rhombic pyroxene is formed from olivine by rearrangement of the independent SiO_4 tetrahedra of the nesosilicate into chains in the inosilicate rhombic pyroxene. During metamorphism this reaction is favored along the olivine-plagioclase interface where disorder and chemical potential are high and is inhibited along the olivine-pyroxene interface where disorder and chemical potential are low. The reaction is initiated by an intergranular film of water which is effective in breaking bonds in olivine. For formation of the second corona minerals, disruption of the plagioclase structure is essential. Inosilicate amphibole is the simplest structural change from plagioclase and is presumably formed in the presence of lower concentrations of water. The formation of nesosilicate garnet requires complete disruption of plagioclase and a change in the co-ordination of Al^{+3}, which probably occurs as a result of greater concentrations of water. Biotite, amphibole, amphibole-spinel, and garnet coronas around iron ore and spinel are also metamorphic and are formed in part by replacement of plagioclase. Regional metamorphism at comparatively high temperatures, under water-deficient conditions and with little shearing stress, gives rise to coronas; the degree of development is partly a function of time. Data and observations are given in support of these conclusions.

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Introduction

Shand (1945) made a critical analysis of observations on the occurrence and origin of rims, sheaths, or mantles of new minerals around olivine, variously described as "coronas," "reaction rims," and "kelyphitic borders." He shows conclusively that the rims are not magmatic, late magmatic, or deuteric. For the development of the rims, factors such as chemical or mineralogical composition, percentage of FeO, or anorthite content of plagioclase of the rock by themselves would not be sufficient. He states that "... the fundamental factor in corona formation may be the instability of olivine under thermal metamorphism, leading to the discharge of iron and magnesium ions which react with plagioclase, the conditions being those of an open system" (Shand, 1945, p. 247). Ellis (1946) thinks that thermal metamorphism induces directed stress resulting in the migration of ions and that consequently the different corona minerals around olivine develop: feldspar is not a source of ions. Gjelsvik (1952) suggests a regional metamorphic origin for the coronites of southern Norway. Huang and Merrit (1954) and Friedman (1955) favor a contact metamorphic origin for the development of coronites. Herz (1951) considers that rims in the Baltimore gabbro are magmatic.

While mapping in parts of south Mirzapur, Uttar Pradesh, and Palamau, Bihar, in India (Survey of India sheets 63 P/4, P/8, L/16, and 64 M/1; scale 1 inch = 1 mile), the writer studied sheets and lenses of metadolerites of ferrogabbroic composition interbanded with Precambrian gneisses, over an area of about 800 square miles. A remarkable feature of these metadolerites is the universal presence of rims around olivine and magnetite. A detailed study was made with special emphasis on the origin of these structures (Murthy, 1955). Similar metadolerites from parts of Bihar, Rajasthan, south India, and North Carolina were studied to supplement observations and verify conclusions.

Present Study

The field work was carried out during the field seasons 1950–1953. Petrographic studies of about 300 thin sections were made in the laboratories of the Geological Survey of India during the years 1953–1955. Refractive-index determinations were made at room temperature (24°C.–27°C.) by the immersion method, using sodium light. The liquids were checked on a Leitz-Jelley refractometer correct to ±0.002. Mineral compositions were computed from Poldervaart's graphs (1950).

The term "rim" is used in this paper in a purely descriptive sense. The term "reaction rim" is restricted to denote rims of undoubted magmatic origin (*e.g.*, Kuno, 1936; 1950; Walker and Poldervaart, 1949, p. 638). The term "corona" is used to describe rims around olivine, iron ore, and spinel which are shown to be metamorphic. Coronas are distinct structurally and mineralogically from reaction rims. Coronites are corona-bearing rocks. In amphibole coronites the second rim is dominantly amphibole (with or without spinel), and in garnet coronites the second rim is dominantly garnet. In the following pages the characteristic features of coronas and coronites are described, and their origin and geological significance are discussed.

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First Corona Around Olivine

In the Indian coronites as in most coronites, the first corona around olivine is invariably rhombic pyroxene, developed by peripheral conversion of olivine, which proceeds from the margin inward. Initially, the prisms of the corona are arranged radially around the olivine core (Fig. 1).

In advanced stages the place of the olivine core is taken by nodular aggregates of haphazardly arranged rhombic pyroxene prisms (Fig. 2).

Coronas may develop around olivines of any composition. In the rocks studied the olivine ranges from Fa_4 to Fa_{74} (Table 1). Whether corona rhombic pyroxene is iron-rich or magnesium-rich depends on the composition of the olivine from which it is derived. (*Cf.* Ramberg and de Vore, 1951, p. 195–196.) Olivines of similar composition may give rise to rhombic pyroxenes of different compositions. For example, olivine Fa_{42} may yield rhombic pyroxene Of_{22} or Of_{28}; olivine Fa_{67-68} may yield rhombic pyroxene Of_{45} or Of_{40} (Table 1). The degree of corona development may vary around different olivine crystals even in the same thin section. Around a single crystal the width of the corona may vary, and parts of the crystal may not show any corona (Fig. 3).

The degree of development of coronas is independent of the size of the olivine crystals.

In most coronites coronas are well developed between olivine and plagioclase but are absent or less well developed between olivine and pyroxene (Figs. 4, 5).

Second Corona Around Olivine

Most coronites show a second corona, whereas a few have more than two coronas. These consist of radiating fibers and prisms of amphibole (Fig. 1), which in many cases have inclusions of spinel (Fig. 2; Santhal Parganas, India, and North Carolina) and/or garnet (Figs. 1, 6; south Mirzapur and Palamau, India, and southern Norway). The second corona is not formed if the first corona is absent.

The second corona is largely contiguous with the first (Figs. 1, 4), although it may not envelop the first (Fig. 1). This led Gjelsvik (1952, p. 100) to consider that the development of corona garnet and rhombic pyroxene is simultaneous. The second corona is formed only when the first is developed, but the second may not form even if the first is well developed (Fig. 7).

The width of the second corona is variable and is not proportionate either to the size of olivine crystals or to the width of the rhombic pyroxene corona which it envelops (Figs. 1, 4).

The observation that the second corona is developed by replacement of plagioclase and not by conversion of the first corona is true of the Indian coronites studied by the writer. The contact between the first and second coronas is in many places sharp and marks the original contact between olivine and plagioclase (Figs. 1, 4, 8).

The second corona is everywhere absent between olivine and pyroxene (Fig. 4). In the development of second corona minerals pyroxene does not seem to contribute (Figs. 4, 9; *cf.* Levin, 1950, p. 540–541).

In many cases multiple twinning of the plagioclase is preserved within the corona amphibole (Fig. 8; Shand, 1945, p. 261). The corona amphibole varies in color and presumably in composition in different coronites. Tremolite, actinolite, and cummingtonite are recorded (Shand, 1945). Spinel inclusions in corona amphibole may be dusty (Fig. 1), granular or vermicular (Fig. 2). Garnet coronas may form symplectites with rhombic pyroxene, recrystallized augite (Figs. 9, 10), and biotite. Most garnet is subhedral toward the plagioclase and anhedral toward pyroxene or amphibole.

Coronas Around Iron Ore and Spinel

In most coronites there are also coronas around iron ore, associated sulfides like pyrite, and green spinel. In most cases the first corona consists of deep-reddish-brown biotite arranged radially around the core of opaque ore. In garnet coronites, the second rim surrounding the biotite rim is generally garnet, and the contact between them is irregular (Fig. 11).

In amphibole coronites, the second rim may consist of amphibole-spinel (Brøgger, 1934, p. 76). In coronites from Santhal Parganas, India, green spinel may be surrounded by coronas of biotite and brown hornblende. As in the coronas

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

FIGURE 1.—MICROGRAPH OF AMPHIBOLE-GARNET CORONITE, SOUTH MIRZAPUR

Note clear olivine (Ol) with radial prisms of corona rhombic pyroxene (Opx) and second corona of amphibole (Amp) locally with inclusions of spinel (Amp-Sp), in a few places garnet (G), specks of recrystallized pyroxene (Repx), and flakes of biotite (Bi). Plagioclase (Pl) is clear. × 40

FIGURE 2.—MICROGRAPH OF AMPHIBOLE-SPINEL CORONITE, SOUTH MIRZAPUR

Note clear olivine (Ol) surrounded by haphazardly oriented corona orthopyroxene (Op) and an outer corona of amphibole (Amp) with inclusions of vermicular spinel (Sp). Peripheral to this corona there are irregular specks of zoisite (Z) within the intensely clouded plagioclase (Pl). Large prisms of primary orthopyroxene (Op^1) are also present. × 23

FIGURE 3.—MICROGRAPH OF CORONITE, SOUTH MIRZAPUR

Note clear olivine (Ol) surrounded partly by corona orthopyroxene (Op). In places corona is absent. Plagioclase (Pl) cores are clouded and vary in intensity in different grains. Interstices (Int) show specks of recrystallized pyroxene. Bi: Biotite. × 40

FIGURE 4.—MICROGRAPH OF AMPHIBOLE CORONITE, SOUTH MIRZAPUR

Note intensely clouded olivine (Ol) surrounded by radial prisms of corona orthopyroxene (Op) and a second corona of fibrous amphibole (Amp) replacing plagioclase (Pl) with clouded cores. Clinopyroxene (Cp) s unclouded. Note the absence of second amphibole corona between olivine and clinopyroxene. × 22

around olivine, there is considerable variation in the degree of development of coronas around iron ore and spinel, which is not dependent on the grain size of the core.

The corona biotite peripherally replaces plagioclase. Amphibole and garnet coronas also develop by replacement of plagioclase (Fig. 11). In the same thin section the amphibole around the corona rhombic pyroxene may be bluish green, whereas the amphibole around magnetite is green. Differences in color may be related to the presence of titanium in the amphibole around iron ore and its absence in the bluish-green amphibole around corona rhombic pyroxene. In some coronites iron ore may show biotite coronas, even where olivine has no coronas (Fig. 12).

Some Characteristic Features of Coronites

General Statement

Detailed studies of thin sections and descriptions in the literature bring out many features common to coronites which are absent in unmetamorphosed mafic igneous rocks of similar bulk composition. These include the phenomena of clouding of primary minerals, recrystallization of pyroxene, and development of garnet.

Clouding Phenomenon

General statement.—The presence of inclusions (specks, hairlike, rodlike) of iron ore or spinel, varying from submicroscopic dimensions to .1 mm or more in diameter, within primary minerals of dolerites such as plagioclase, pyroxene, and olivine is termed "clouding." The significance of clouding especially in plagioclase has been discussed by Poldervaart and Gilkey (1954). Clouding is different from alteration, which involves the development of new minerals from host minerals.

Plagioclase.—The plagioclase of almost all coronites shows gray or brown clouding (Figs. 2, 3, 4, 5, 8, 10; *cf.*, Lacroix, 1889; Sederholm, 1916; Teale, 1923; Brøgger, 1934; Shand, 1945; Gjelsvik, 1952). This is due to the presence of ultramicroscopic or larger specks of iron ore (MacGregor, 1931, p. 524), spinel (Shand, 1945, p. 261), clinopyroxene (Tilley, 1921, p. 117; also in south Mirzapur and Palamau), or mica-like minerals (Gjelsvik, 1952, p. 49). X-ray fluorescence and chemical analysis show that clouding in plagioclase of North Carolina metadolerites is due to formation of Fe-rich rhombic pyroxene (Poldervaart, 1955, personal communication). The specks may be apparently unoriented (Fig. 4), in many cases arranged along twin composition planes (Figs. 3, 8, 10) and in a few cases along zone boundaries in rhythmically zoned plagioclase (Fig. 5). Poldervaart and Gilkey (1954, p. 88) advanced convincing arguments in favor of a metamorphic origin (regional or thermal) for the intense clouding of plagioclase.

Figure 5.—Micrograph of Coronite, South Mirzapur

Note that olivine is completely replaced by orthopyroxene (Op). Core shows specks of iron ore. Intensely clouded clinopyroxene (Cp) is in places clear in the periphery bordering the interstices (Int) which contain specks and vermicules of recrystallized pyroxene. Plagioclase (Pl) is intensely clouded, and in one crystal the clouding material is arranged in zones. Iron ore (I) is in places surrounded by biotite (Bi) corona. Ap: Apatite. × 44

Clinopyroxene.—The clinopyroxene of coronites may also show clouding (Figs. 5, 6, 9, 11). In the earliest stages, submicroscopic dust in many cases is concentrated along the periphery (Fig. 12), and such crystals show distinct pleochroism or absorption from pink to brown; the intensity varies with the concentration of inclusions. (*Cf.* Krishnan, 1926, p. 403; Temperley, 1938, p. 32.) In coronites from south Mirzapur, Palamau, India (Figs. 5, 9), Adirondacks (Buddington, 1939), and southern Norway (Gjelsvik, 1952), innumerable hairlike, rodlike, or granular inclusions of iron ore in many cases render the primary pyroxene almost opaque. Clouding of pyroxene which accom-

TABLE 1.—OPTICAL PROPERTIES AND ESTIMATED COMPOSITIONS OF OLIVINE AND RHOMBIC PYROXENE
(Compositions after Poldervaart, 1950)

Coronite type	Olivine		Rhombic pyroxene	
	Optical data	Composition	Optical data	Composition
Amphibole spinel				
1	$\beta = 1.670$	Fa_{9}		
2	$\beta = 1.717$	Fa_{31}		
3	$\beta = 1.660$–1.711	Fa_{4}-Fa_{28}	$\gamma = 1.676$–1.701	Of_{5}-Of_{27}
4	$\beta = 1.662$–1.712	Fa_{5}-Fa_{29}	$\gamma = 1.700$	Of_{26}
5	$\alpha = 1.715$	Fa_{42}	$\gamma = 1.696$	Of_{22}
6	$\alpha = 1.676$	Fa_{21}	$\gamma = 1.686$	Of_{14}
7	$\alpha = 1.702$	Fa_{35}	$\gamma = 1.700$	Of_{26}
8	$\alpha = 1.709$	Fa_{39}	$\gamma = 1.698$	Of_{24}
9	$\alpha = 1.687$	Fa_{27}	$\gamma = 1.693$	Of_{20}
10	$\alpha = 1.689$	Fa_{28}	$\gamma = 1.689$	Of_{18}
11	$\alpha = 1.679$	Fa_{23}	$\gamma = 1.676$	Of_{5}
12	$\alpha = 1.695$	Fa_{31}	$\gamma = 1.695$	Of_{22}
13	$\alpha = 1.714$	Fa_{42}	$\gamma = 1.703$	Of_{28}
Clinopyroxene spinel				
14	n = 1.691 $2V_{\alpha} = 87°$	Fa_{20}	$2V_{\alpha} = 82°$	Of_{15}
Hypersthene spinel				
15	$2V(-) = 84°$–$86°$	Fa_{33}-Fa_{38}	$2V(-) = 72°$	Of_{24}
Amphibole (spinel)-garnet				
16	$\alpha = 1.740$	Fa_{55}	$\gamma = 1.713$	Of_{37}
Garnet				
17	$\gamma = 1.732$–1.762	Fa_{29}-Fa_{48}	$\gamma = 1.690$–1.700	Of_{17}-Of_{26}
18	$\alpha = 1.760$	Fa_{66}	$\gamma = 1.713$	Of_{37}
19	$\alpha = 1.762$	Fa_{67}	$\gamma = 1.723$	Of_{45}
20	$\alpha = 1.764$	Fa_{69}	$\gamma = 1.716$	Of_{40}
Garnet, and some amphibole (no spinel)				
21	$\alpha = 1.753$	Fa_{62}	$\gamma = 1.717$	Of_{40}
Coronite with only rhombic pyroxene				
22	$\alpha = 1.774$	Fa_{74}	$\gamma = 1.730$	Of_{52}
Olivine metadolerite with no coronas				
23	1.749	Fa_{61}		

1. Olivine gabbro, Sulitelma (Vogt, 1927, p. 287).
2. Basic olivine gabbro, Sulitelma (Vogt, 1927, p. 287).
3. Baltimore gabbro (Herz, 1951, p. 985, 987).
4. Wichita troctolite (Huang and Merrit, 1954, p. 535, 557).
5. Olivine metadolerite (473), Rajasthan, India (specimen loaned by Karunakaran).
6. Olivine metadolerite (1102), south Mirzapur, India.
7. Olivine metadolerite (7181), south Mirzapur, India.
8. Olivine metadolerite (170), south Mirzapur, India.
9. Olivine metadolerite (1138), south Mirzapur, India.
10. Olivine metadolerite (C.S.R.4), Santhal Parganas, Bihar, India (specimen loaned by Raja Rao).
11. Olivine metadolerite (656), Manbhum, Bihar, India (specimen loaned by Mahadevan).
12. Olivine metadolerite (804), Manbhum, Bihar, India (specimen loaned by Mahadevan).
13. Olivine metadolerite, North Carolina (specimen loaned by Poldervaart).
14. Picrite, Caribou Lake intrusive, McConkey Township, Ontario, Canada (Friedman, personal communication).

15. Troctolite, Eastern Quebec, Canada (Shand, 1945, p. 260).
16. Olivine metadolerite (296), south Mirzapur, India.
17. Olivine metagabbro, southern Norway (Gjelsvik, 1952, p. 72, 76).
18. Olivine metadolerite (13), south Mirzapur, India.
19. Olivine metadolerite (480), south Mirzapur, India.
20. Olivine metadolerite (998), south Mirzapur, India.
21. Olivine metadolerite (439), south Mirzapur, India.
22. Olivine metadolerite, Santhal Parganas, Bihar, India (specimen loaned by Raja Rao).
23. Olivine metadolerite (1024), south Mirzapur, India.

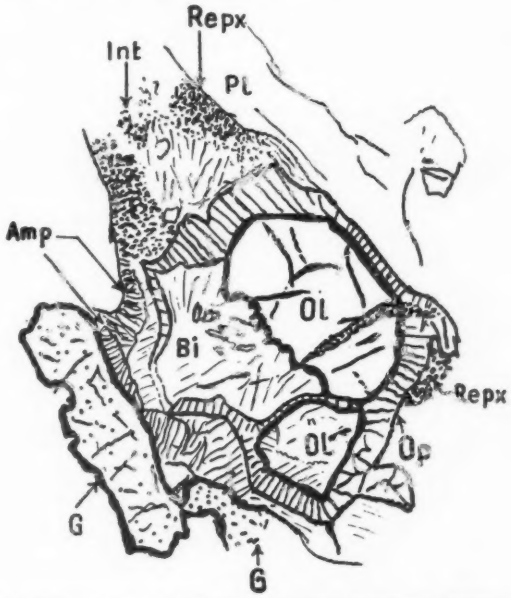

FIGURE 6.—MICROGRAPH OF AMPHIBOLE-GARNET CORONITE, SOUTH MIRZAPUR

Clear olivine (Ol) is surrounded by orthopyroxene (Op) corona and a second corona of garnet (G) and amphibole (Amp). Part of the olivine is replaced by biotite (Bi) flakes; the first corona of orthopyroxene is only partly preserved. Interstices (Int) show vermicules of recrystallized pyroxene (Repx). × 80

panies clouding of plagioclase is also caused by metamorphism. (*Cf.*, MacGregor, 1931, p. 534.)

Rhombic pyroxene.—Primary rhombic pyroxene of coronites may also show submicroscopic dusty particles and plates of iron ore (*e.g.*, south Mirzapur and southern Norway). The corona rhombic pyroxene is everywhere unclouded.

Olivine.—The olivine of coronites may also show clouding which may be intense, as in the south India metanorites (Fig. 4; *cf.* Holland, 1897, p. 24), and also in two coronites from Manbhum, Bihar, India (collected by T. M. Mahadevan). Most of the inclusions (minute specks of ? magnetite) are oriented parallel to the *c* axis, and such crystals may show absorption from brown to dark gray. In the south India metanorites plagioclase and olivine show clouding, but the pyroxenes do not (Fig. 4). In the south Mirzapur and Palamau coronites, the pyroxenes are intensely clouded, but the olivines do not show clouding. Clouded olivines have been noted by Hills (1900, p. 534), MacGregor (1931, p. 534), and Poldervaart and Gilkey (1954, p. 78).

Recrystallization of Pyroxene

In the south Mirzapur and Palamau coronites specks, vermicules, and granular aggregates of colorless or pale-green recrystallized clinopyroxene occur within interstices (Figs. 1, 5, 6, 7, 9, 10). Recrystallized pyroxene is considered to be characteristic of metabasaltic

FIGURE 7

FIGURE 8

FIGURE 9

FIGURE 10

FIGURE 7.—MICROGRAPH OF CORONITE, NANDALALPUR, MANBHUM, BIHAR

Clear olivine (Ol) is surrounded by orthopyroxene (O.Px) corona. Note absence of second coronas. Dark patches of iron ore are surrounded by biotite (Bi) corona. Clinopyroxene (Cp) is intensely clouded. Plagioclase (Pl) is clouded in the cores. Recrystallized pyroxene (RePx) occurs in the interstices. Ap: Apatite. × 22

FIGURE 8.—MICROGRAPH OF AMPHIBOLE CORONITE, NORTH CAROLINA

Clear olivine (Ol) is surrounded by orthopyroxene (Op) corona and a second corona of amphibole (Amp) prisms which replace plagioclase (Pl). Note intense clouding of plagioclase along twin composition planes, preserved in the later amphibole corona. Clinopyroxene (Cp) is intensely clouded. Bi: Biotite. × 40

FIGURE 9.—MICROGRAPH OF GARNET CORONITE, SOUTH MIRZAPUR

Clear olivine (Ol) with orthopyroxene (Op) and garnet (G) corona. Iron ore (I) is surrounded by biotite (Bi) and garnet corona which encloses vermicules of recrystallized pyroxene. Biotite replaces part of olivine which is partly altered to bowlingite (Bow); part of the orthopyroxene corona is preserved. One of the olivine crystals shows a narrow orthopyroxene corona along the contact with clinopyroxene (Cp) which is intensely clouded. Plagioclase (Pl) shows clouding of the cores. Ap: Apatite. × 44

FIGURE 10.—MICROGRAPH OF GARNET CORONITE, SOUTH MIRZAPUR

Iron ore (I) is surrounded by biotite (Bi) and hornblende (H) coronas. Nodular aggregates of orthopyroxene (Op) represent original olivine. Garnet (G) is later than the intense clouding of plagioclase (Pl) which it replaces. Garnet also forms symplectites with vermicules of recrystallized pyroxene (RePx) occurring in interstices (Int). × 38

rocks showing "abnormal granulitic" trend of metamorphism (Poldervaart, 1953a, p. 264) and has been noted by Stillwell (1918), Tilley (1921), Temperley (1938), and Sutton and Watson (1951b).

nites, the primary pyroxene is intensely clouded, but the recrystallized pyroxene is not. In slices of these coronites, the mineral that clouds the plagioclase appears to be similar to the recrystallized pyroxene. (*Cf.*, Poldervaart and Gilkey,

FIGURE 11

FIGURE 12

FIGURE 11.—MICROGRAPH OF GARNET CORONITE, SOUTH MIRZAPUR

Iron ore (I) is surrounded by biotite (Bi) corona and partly by garnet (G) and hornblende (H) coronas which develop by replacement of plagioclase (Pl) showing only mild clouding. Clinopyroxene (Cp) shows intense clouding. Corona orthopyroxene (Op) is seen. Recrystallized pyroxene specks and aggregates and apatite (Ap) are seen in the interstices (Int). × 37

FIGURE 12.—MICROGRAPH OF METADOLERITE, SOUTH MIRZAPUR

Note the absence of coronas around clear olivine (Ol). Narrow biotite (Bi) corona around iron ore (i). Clinopyroxene (Cp) shows mild clouding along the periphery. Plagioclase (Pl) is clear. Specks of recrystallized pyroxene (RePx) occur in the interstices. × 43

Development of Garnet

The formation of garnet coronas in coronites is also significant. Garnet is not formed during crystallization of basaltic magma.

INTERRELATIONSHIP BETWEEN CORONA FORMATION AND METAMORPHIC PHENOMENA SUCH AS CLOUDING OF MINERALS

From petrographic studies of coronites the following conclusions may be derived:

(1) Clouding of plagioclase is earlier than corona formation. In the coronites from eastern Quebec (Poldervaart and Gilkey, 1954, p. 84) and North Carolina, the plagioclase shows clouding especially along twin composition planes recognizable as parallel planes of dusty inclusions within the hornblende coronas (Fig. 8). In the Mirzapur coronites, corona garnet is later than clouding of plagioclase (Figs. 9, 10).

(2) In the south Mirzapur and Palamau coronites, 1954, p. 87.) Clouding of pyroxene appears to be slightly later or contemporaneous with recrystallization of pyroxene (Fig. 9). Recrystallization of pyroxene is earlier than the formation of corona garnet which in some cases shows vermicular inclusions of pyroxene (Figs. 9, 10).

(3) In the south India metanorites clouding of olivine is earlier than the formation of the first corona (Fig. 4).

(4) In many cases biotite is later than corona rhombic pyroxene (Figs. 6, 9).

Although all the relations are not observed everywhere, it is reasonable to conclude that metamorphism gives rise to clouding, recrystallization, and development of garnet and is also responsible for the development of coronas around olivine and iron ore.

ORIGIN OF CORONAS

First Corona Around Olivine

In the formation of the first corona from olivine, the independent SiO_4 tetrahedra of the

nesosilicate are converted into chains of linked SiO_4 tetrahedra, each sharing two oxygen atoms in the inosilicate rhombic pyroxene. (*Cf.* Poldervaart, 1953b, p. 493.) This requires the breaking of the electrostatic bonds between the two cations (Fe^{+2} and/or Mg^{+2}) and the two oxygen atoms of the silicon tetrahedra, probably achieved by the action of water which is a powerful agent in weakening bonds (Mason, 1952, p. 223). In this process two oxygen atoms are released from SiO_4 and presumably form OH^{-1} ions. The trend of the reaction may be represented schematically by the following equation:

$$(Mg,Fe)_2SiO_4 + H_2O \rightarrow (Mg,Fe)SiO_3 + Fe^{+2} + Mg^{+2} + 2OH^{-1}$$

Olivine + water → Rhombic pyroxene + ions

Petrographic studies indicate that the development of rhombic pyroxene is favored along the olivine-plagioclase interface but inhibited along the olivine-pyroxene interface where coronas are either absent or only poorly developed. Along the olivine-plagioclase interface, the nesosilicate olivine with four active oxygen atoms is adjacent to the tectosilicate plagioclase in which oxygen atoms are inert since they are shared by neighboring SiO_4 and AlO_4 tetrahedra in a three-dimensional network. During metamorphism both disorder and chemical potential are high along the interface, since the two adjacent minerals are crystallochemically highly contrasted. The small amount of water permeating along grain boundaries reacts with the "active" oxygen atoms of the olivine structure, and the rhombic pyroxene corona is formed; the plagioclase, however, remains stable. The reaction once started along the olivine-plagioclase interface can then proceed along the interface between differently oriented olivine grains in a glomeroaggregate. The rarity of coronas between olivine and primary pyroxene can also be explained on the same basis. Both olivine and primary pyroxene have Fe^{+2} and Mg^{+2} ions, and therefore chemical potential and disorder are low at the interface of the grains, which inhibits the catalytic action of water on olivine. The commonly noted variation in the degree of corona formation and the lack of correspondence between the size of the olivine grains and the width of the coronas are to be expected, since reactions take place along interfaces where chemical potential, disorder, and water concentrations cannot be uniform. The variable composition of the rhombic pyroxene (Table 1), can also be explained on the basis of free energy which results in different rates of diffusion of the cations.

Second Corona Around Olivine

Studies of coronites establish that the formation of amphibole (plus spinel) and/or garnet takes place by the replacement of the plagioclase adjacent to the rhombic pyroxene corona. The development of these minerals from plagioclase requires additions of Fe^{+2} and Mg^{+2} ions which are released from olivine during the formation of rhombic pyroxene coronas.

Coronites in a particular area may develop either amphibole (Santhal Parganas, India) or garnet coronas (southern Norway). Both types of coronas may develop in the same rock. (*Cf.* Brøgger, 1934, p. 24.) Although rocks with magnesian olivines apparently tend to develop amphibole coronas, and iron-rich olivines preferentially develop garnet coronas, in some rocks both minerals may form from olivines of the same composition (Fig. 1; Table 1).

Since there are unfortunately little or no data on the composition of corona amphiboles, analyses of amphiboles in metagabbros given by Buddington (1952) are compared in Table 2 with the few available chemical analyses of corona garnets (Shand, 1945; Levin, 1950; Buddington, 1952).

The table brings out the contrast in chemical composition between the two minerals. Corona garnets are rich in (Fe,Mn)O and Al_2O_3, whereas the amphiboles are rich in MgO and CaO.

The formation of the second corona minerals involves a change from the tectosilicate structure of plagioclase to the inosilicate structure of amphibole or the nesosilicate structure of garnet. The problem, therefore, is how minerals of such divergent composition and structure are formed by reaction of plagioclase with migrating Fe^{+2} and Mg^{+2} ions.

For the formation of amphibole, the three-dimensional network of SiO_4 and AlO_4 tetrahedra of the plagioclase must be converted into parallel chains of SiO_4 tetrahedra. This involves breaking of the bonding between AlO_4 and SiO_4 tetrahedra of the feldspar, and part of the Al^{+3} must assume 6 co-ordination. In this process also, water probably plays an active role. Since all the oxygen atoms in the plagioclase structure are mutually shared and therefore inert, the bonds are stronger than in the olivine structure in which all the oxygen atoms are unshared and therefore "active." It is therefore probable

that greater concentrations of water are necessary in disrupting the bonds in plagioclase than in olivine. The disruption is likely to take place with relative ease between the Al-O-Si links. The bonds in the albite structure are likely to be less strong than in the anorthite structure, since only a quarter of the tetrahedra are of AlO_4, and therefore most of the oxygen atoms will be shared by neighboring Si^{+4} ions. In the anorthite structure half the oxygen atoms will be shared by neighboring Al^{+3} and Si^{+4} ions. (*Cf.* Loewenstein, 1954, p. 93.) According to Bragg (1937, p. 234), in orthoclase the bonding that links the chains is weak. This is likely to be true of plagioclase also, and probably these links are comparatively easily broken by the catalytic action of water. Thus parallel silicate chains requisite for the formation of corona amphibole are the most easily formed from plagioclase. Once the essential structural units are formed, amphibole, which is apparently the stable mineral under the p-t-H_2O conditions, is formed with Ca^{+2} ions released from anorthite and Mg^{+2} and Fe^{+2} migrating from olivine. Since OH^{-1} ions are an essential constituent of amphibole, part of the water is taken up by the structure. The excess Al^{+3} released from the plagioclase during this process may either migrate or combine with Mg^{+2} and Fe^{+2} to form spinel, which in many cases is seen as inclusions within corona amphibole.

The formation of corona garnet involves complete disruption of the plagioclase, possibly when higher concentrations of water are available in the system. All the bonds between the SiO_4 and AlO_4 units are broken. Al^{+3} is in 4 co-ordination in plagioclase but in garnet has an 8 co-ordination. Formation of garnet, therefore, also requires a change in the co-ordination of Al^{+3}. It is suggested that this is brought about by temporary substitution of O^{-2} by OH^{-1}; Al^{+3} has preferred octahedral co-ordination with OH^{-1}. (*Cf.* Hauser, 1955, p. 11.) Since Fe^{+2} ions prefer a nesosilicate structure (*cf.* Ramberg, 1952, p. 340) corona garnets are enriched in this ion relative to Mg^{+2} which prefers linked structures.

If the foregoing reasoning is correct, the development of both amphibole and garnet in the same coronite may be explained as a result of progressive increase in water concentrations; where only garnet is formed, rapid increase in water concentrations is presumably indicated. The hypothesis also explains why the type of corona formed is not related either to the bulk composition of the rock or to the composition of the olivine from which the coronas are formed. The type of corona mineral formed seems to depend on the concentrations of water, which control the degree and type of disruption of the plagioclase structure.

The thesis developed here also explains the absence of the second corona even where the first corona is well developed (Santhal Parganas; Fig. 7). During metamorphism of this rock, apparently only small amounts of water were available; they were sufficient only for the formation of the first corona from olivine but not sufficient to disrupt the plagioclase structure.

TABLE 2.—CHEMICAL COMPOSITION OF AMPHIBOLES IN METAGABBROS AND CORONA GARNETS

(In per cent)

	Amphibole	Garnet
CaO	28–33	6–7
MgO	39–52	11–12
(Fe, Mn)O	9–14	17–20
Al_2O_3	11–15	20–22

Coronas around Iron Ore and Spinel

The commonly found coronas of biotite, amphibole, or garnet around iron ore and spinel are formed mainly by contributions from the minerals around which they develop and partly from plagioclase which they replace. Fe^{+2} and Fe^{+3} ions are supplied by iron ore, and Mg^{+2} ions are contributed from olivine after the formation of the first rhombic pyroxene corona. In some coronites, however, iron ore may show coronas, even in the absence of coronas around olivine (Fig. 12). Here Mg^{+2} ions may be supplied by the magnetite; many analyses of magnetites show MgO. (*Cf.* Palache *et al.*, 1944, p. 701.) Al^{+3}, Si^{+4}, Na^{+1}, and K^{+1} are available in plagioclase and potassium feldspar if present (in south Mirzapur, in anorthoclase of the interstices). In many cases biotite replaces the olivine core, whereas the rhombic pyroxene corona is retained (Figs. 6, 9). This suggests that at least part of the biotite is later than the rhombic pyroxene corona. This biotite is presumably formed by migration of Fe^{+3}, K^{+1}, and Al^{+3} ions from outside the olivine structure. The nesosilicate olivine is apparently more readily converted into biotite than the inosilicate rhombic pyroxene. Neither amphibole nor garnet replaces olivine.

Spinel, like magnetite, seems to be crystallochemically ideal for the development of coronas.

As in the formation of coronas around olivine, during the formation of coronas around iron ore and spinel also, water presumably plays an active role in breaking bonds. The formation of these coronas also is due to diffusion and substitution.

Petrology of Coronites

At low grades of metamorphism (low temperatures and low hydrostatic pressure) no coronas are formed in olivine metadolerites. At low temperatures, possibly aided by shearing stress which results in the development of fractures, these rocks tend to acquire water (*cf.* Thompson, 1955, p. 98–99) from the country rocks, and metamorphism usually results in the formation of rocks of the greenstone, epidote-amphibolite, and amphibolite facies, and predominantly hydrous minerals such as serpentine, epidote, zoisite, chlorite, and amphibole are developed. Original igneous structures are invariably destroyed.

Coronites, however, characteristically preserve original igneous structures. The primary minerals retain their form though not their composition. Buddington (1952, p. 49–52) gives examples of metagabbros with apparent relics of plagioclase which have become more sodic, based on comparisons of the actual and normative anorthite contents of the mineral. Similarly it can be shown that the olivines of the south Mirzapur and Palamau coronites have become iron-rich without change in form, during metamorphism; the actual composition of the olivine in two analyzed specimens is Fa_{66}, whereas the norm gives a composition of Fa_{56}. Development of new minerals is limited in extent and degree.

In the formation of coronites by metamorphism, the prime process is ionic diffusion and substitution. These processes depend on two important factors, *i.e.*, elevated temperatures and the presence of intergranular water vapor.

The importance of elevated temperatures in diffusion during metamorphism has been stressed by Buerger (1948) and Turner and Verhoogen (1951, p. 408). Elevated temperatures result in disorder, especially along crystal boundaries. But such disorder and the consequent conditions suited for diffusion may not by themselves bring about reactions leading to corona formation.

In ionic diffusion and substitution, the writer suggests that water vapor plays an important role in breaking bonding in minerals, in liberating cations held tightly in crystal structures, in hastening diffusion, and in the development of new minerals by aiding the formation of new structural units. (*Cf.* Turner and Verhoogen, 1951, p. 409). In the development of clouded plagioclase commonly noted in coronites, Poldervaart and Gilkey (1954) considered water vapor essential for diffusion of ions.

The water content of basaltic rocks is low, usually around 1 per cent (Green and Poldervaart, 1955, p. 179, 182). The chemical analyses of two specimens of coronites from south Mirzapur and Palamau show only 0.43 per cent water. Igneous rocks in general have low porosity and permeability (Yoder, 1955, p. 518). Possibly the small amount of "potential" water present in the dolerites as OH^{-1}, H_2O, and H_3O^{+1} can become available for reactions, during metamorphism when the rocks are heated. For the formation of coronas, evidently only small amounts of water are required. When excess water conditions prevail plagioclase recrystallizes, pyroxenes are converted into amphiboles, and the basaltic rocks are converted into amphibolites (*e.g.*, Santhal Parganas, India; and southern Norway: Gjelsvik, 1952) Yoder (1955, p. 519–520) has shown that the development of the most hydrous assemblage from an elongated body of basic igneous rock need not mean change in p-t conditions but only changes in water concentrations.

In summary, the development of coronas and coronites takes place during regional metamorphism at elevated temperatures when water-deficient conditions prevail. The reactions leading to corona formation are slow and take place by diffusion processes by ionic substitution and subtraction along the interfaces of mineral grains in the solid state in an essentially closed system. The prerequisites for such processes are the presence of high chemical potential and disorder along grain boundaries, which are determined by the crystallochemical nature of the minerals and prevailing high temperatures. But the initiation and continuation of reactions leading to the development of coronas depend on the availability of water vapor diffusing along intergranular boundaries. The relative stability of the minerals along an interface depends on the degree of ease with which water can react with minerals, which is of course related to the crystallochemical nature of the minerals and the prevailing disorder. With small amounts of water, olivine is unstable, but plagioclase is stable; only rhombic pyroxene coronas are formed. With increasing amounts of water, plagioclase is also unstable, and amphibole

coronas may be formed. With further increase in water, garnet coronas are formed.

Coronites are well developed in plutonic terrains (*Cf.* Sutton and Watson, 1951b, p. 34) and are absent in predominantly schistose areas, which suggests that for the development of coronas hydrostatic pressure must be more important than shearing stress.

In south Mirzapur and Palamau corona development is ubiquitous in the individual sheets and lenses; a specimen taken from any portion of the metadolerite shows coronas. Both in the Indian occurrences and elsewhere (southern Norway, Sweden, Adirondacks, North Carolina) coronites are invariably developed on a regional scale, *i.e.*, by regional metamorphism. Such metamorphism, postdating the olivine-bearing mafic intrusives, has been established by Read (1923) and Sutton and Watson (1951a) in Scotland, by Buddington (1939; 1952) in the Adirondacks, and by Kulp and Poldervaart (1956, p. 393–403) in North Carolina. Later pegmatite and quartzofeldspathic intrusives have been noted by the writer in south Mirzapur, by T. M. Mahadevan (Personal communication) in Manbhum, Bihar, by C. Karunakaran (Personal communication) in Rajasthan, India, by Huang and Merrit (1954) in Oklahoma, and by Friedman (1955) in Canada. However, it is not correct to infer that the pegmatites are responsible for the development of coronas. (*Cf.* Friedman, 1955, p. 595.) The later intrusives are manifestations of metamorphic processes which have operated after intrusion of the mafic rocks (a younger metamorphic cycle).

In terrains where coronites are well developed, it may not always be possible to recognize postdolerite intrusives, the absence of which does not invalidate a metamorphic origin for coronites. On the basis of the evidence presented in this paper, the occurrence of coronites with their characteristic metamorphic features can be taken as evidence of a water-deficient metamorphism at elevated temperatures, after the intrusion of the mafic rocks.

Temperature of Corona Formation

Coronites belong to the granulite facies of metamorphism which according to Rosenquist (1952) develops from 500° to 900° C. for depths of 11 and 6 miles respectively. In a garnet amphibolite at the Barton mine where rocks belong to the granulite facies (*cf.* Buddington, 1952, p. 83) the decrepitation temperature of pyralmandite garnet and hornblende is given by Smith (1952, p. 473) to be 613° ± 20°C. and 603° ± 20°C. respectively. He suggests that the rocks were formed at about these temperatures (corrected for the effect of pressure during recrystallization). For the formation of the Adirondack granulites Buddington and others (1955, p. 503) suggest temperatures between 500° and 700°C.

Experimental studies by Yoder (1952; 1955) lead him to conclude that iron garnet (almandite) is stable below 785°C. in the absence of water. Yoder's (1952, p. 585) studies show that enstatite grows at about 660°C. from a glass of proper composition. Enstatite is also stable in the water-deficient region below this temperature, which is likely to be lowered for iron-rich hypersthene.

Poldervaart (1955, personal communication) states that clouding of plagioclase probably takes place between 500° and 600°C. Corona formation is later than clouding. From these observations the writer concludes that the formation of rhombic pyroxene and garnet presumably take place between 550° and 600°C. Amphibole and amphibole-spinel and biotite coronas probably also form at about the same or slightly lower temperatures.

Time Factor

Poldervaart and Gilkey (1954, p. 82) note that time plays a significant role in the degree of development of clouding. Assuming that a metamorphic cycle is $\sim 10^6$ years (Yoder, 1955, p. 513) it is evident from the small degree of mineralogical changes in coronites that metamorphic response under water-deficient conditions is slow. The degree of corona formation evidently depends on the time during which the dolerites were subjected to optimum conditions necessary for corona formation.

Contrast between Magmatic Reaction Rims and Metamorphic Coronas

In the following paragraphs, the characteristics of magmatic reaction rims in unmetamorphosed olivine-bearing igneous rocks, mainly based on data from the literature (Kuno, 1936; 1950; Walker and Poldervaart, 1949), are compared with those of coronas in coronites, which shows that the two are distinct structurally and mineralogically, as a result of their contrasted modes of origin.

(1) Reaction rims are common in volcanic andesites and tholeiitic basalts around earlier phenocrysts of magnesian olivines. Iron-rich olivines have no reaction rims, since they do

not bear a reaction relation to rhombic pyroxene (Bowen and Shairer, 1935). In hypabyssal dolerites plates of reaction minerals (enstatite and pigeonite) in some cases enclose cores of olivine (Walker and Poldervaart, 1949, p. 638).

Coronas can develop in any olivine-bearing rock and olivines of any composition.

(2) The width of reaction rims is primarily a function of grain size of the olivine. Smaller crystals are completely converted to pyroxene.

Coronas may develop around all grains of olivine and magnetite crystals which are in contact with plagioclase, irrespective of their size. The degree of reaction depends on the concentrations of water.

(3) The first reaction mineral may be rhombic pyroxene or pigeonite or both, depending on the phases separating from the melt at the time of reaction.

The first corona mineral is invariably rhombic pyroxene.

(4) Most primary rhombic pyroxene of dolerites occurs in long prisms; corona rhombic pyroxene prisms are invariably short.

(5) Reaction minerals are arranged tangentially around olivine cores; corona minerals are arranged radially around olivine, iron ore, or spinel cores.

(6) In magmatic reactions the second and other rims are formed by reaction of olivine with the melt. No amphibole-spinel rims are noted.

In coronites the second corona minerals are formed by replacement of the plagioclase. The second corona mineral may be amphibole, amphibole-spinel, and/or garnet.

(7) Rocks in which reaction rims are well developed do not show clouding of plagioclase, pyroxene, or olivine or recrystallization of primary minerals.

Most coronites have clouded plagioclase and may have clouded pyroxenes and olivines. In many cases primary pyroxene is recrystallized.

(8) In rocks with reaction rims, garnet is absent; coronites may show garnet, which in most cases forms the second corona.

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SHALLOW SUBMERGED MARINE TERRACES OF SOUTHERN CALIFORNIA

BY K. O. EMERY

ABSTRACT

Profiles across the mainland shelf, island shelves, and bank tops of southern California show five separate flattenings that are interpreted as erosional marine terraces cut during times of low sea level of the Pleistocene Epoch. Similar terraces have been discovered recently in widely separated parts of the world; this supports the interpretation of their relationship to eustatic changes of sea level. A correlation diagram of the terraces of southern California shows that each terrace is deeper around offshore islands and banks than off the mainland; this is attributed to regional warping that from other evidence is believed to have begun in Late Miocene time. The warping indicated by the deepest terrace, at the shelf edge, amounts to about 160 feet per 100 miles.

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INTRODUCTION

For many years emerged marine terraces have been described and credited to eustatic rises of sea level during interglacial ages of the Pleistocene Epoch (Stearns, 1935; Zeuner, 1945, p. 225–252; MacNeil, 1949). The writer, and undoubtedly many others, believed that investigations should be directed toward learning the nature and distribution of the related submerged terraces that must have been formed during intervening glacial ages of low sea level. As long ago as 1900 Smith published profiles of the shelf off southern California, but only after the much later development of acoustic sounding equipment could small variations in the levels of the shelves be detected. Using such equipment Norris (1951), reported finding several submerged terraces off San Nicolas Island in southern California. During the fol-

lowing year the writer (Emery, in press) mapped several submerged terraces that fringe the island of Guam. These results led to the beginning of a general survey off southern California in March 1954 to learn how consistent and widespread are the terraces. At first, sounding lines were run only when the University of Southern California's research ship Velero IV was leaving anchorages en route to do other work. When preliminary results showed that the terraces might reveal something of the late geological history of the region, several special cruises totaling about 11 days were set up for the work. With the last of these cruises the survey was completed late in 1956.

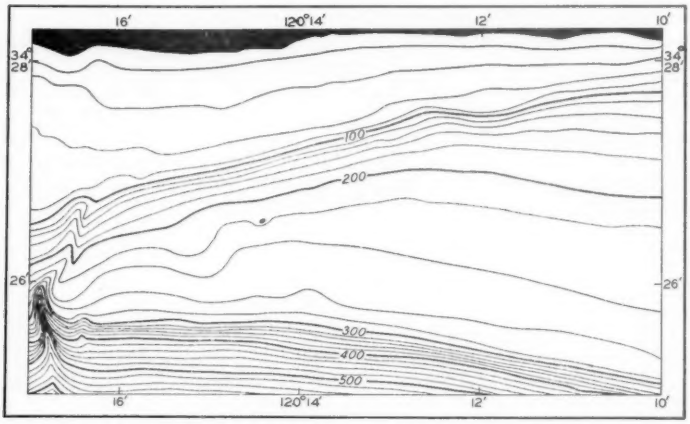

FIGURE 1.—CONTOUR MAP OF SMALL AREA OF MAINLAND SHELF AT PROFILE 2 NEAR POINT CONCEPTION

Two terraces are shown with outer edges at depths of about 70 and 300 feet, respectively. Data are too scanty to reveal other smaller terraces. Contours are from soundings on U. S. Coast and Geodetic Survey smooth sheets 5625 and 5626.

The continental shelf has been defined by the International Committee on Nomenclature of Ocean Bottom Features (Wiseman and Ovey, 1953) as,

"The zone around the continent, extending from the low-water line to the depth at which there is a marked increase of slope to greater depth. Where this increase occurs the term shelf edge is appropriate. Conventionally its edge is taken at 100 fathoms (or 200 meters) but instances are known where the increase of slope occurs at more than 200 or less than 65 fathoms. When the zone below the low water line is highly irregular, and includes depths well in excess of those typical of continental shelves, the term Continental Borderland is appropriate."

Southern California is the type locality of the continental borderland, which consists of many basins and troughs separated by islands and banks (Shepard and Emery, 1941). Nowhere in southern California is the continental shelf more than 15 miles wide, and it exceeds 7.5 miles at only four places. Island shelves of the region are closely related to the continental shelf along the mainland, and both are included by the general term shelf as used in this paper. The terraces that are superimposed atop the shelves should not be confused with the term continental terrace used by Daly (1942) and Dietz (1952), because this term includes both continental shelf and continental slope and thus designates a much broader structure than these small features.

Appreciation is due Captain Allan Hancock for making available ship time, Officers F. C. Ziesenhenne and W. C. Phalen for their careful navigation, and Drs. R. P. Sharp of California Institute of Technology, W. C. Putnam of University of California at Los Angeles, F. P. Shepard, Tj. van Andel, and M. N. Bramlette of Scripps Institution of Oceanography, M. L. Natland of Richfield Oil Corporation, and O. L. Bandy of University of Southern California for fruitful discussions of the terraces and their possible significance. Thanks also are expressed to Drs. W. P. Woodring of the U. S. Geological Survey and Norman D. Newell of the American Museum of Natural History for critical reading of the manuscript.

Survey Methods

The original detail sheets made by the U. S. Coast and Geodetic Survey contain many soundings which when contoured commonly reveal the presence of two or more wide terraces on the continental shelf (Fig. 1). The contours show that these terraces continue for miles along the coast and thus that they are not merely the tops of stray slump blocks. However, in most areas the soundings are not sufficiently dense to reveal the smaller terraces. Individual profiles made up of closely spaced soundings were more useful than contours for showing small irregularities, but uncertainties of correlation from profile to profile arise.

Sounding lines were run at right angles to the coast (or to the bottom contours) at spacings of 1–16 miles. The lines started as close to the shore line as the depth of water, maneuvering room, and sea conditions permitted, and most of them ended at a depth of 700 feet, well beyond the shelf edge. Lengths of the lines ranged from 300 to 12,000 yards. Ship speed was kept at half—about 7 knots. All soundings were read visually to the closest foot from a small 600-foot scale echo sounder, a Raytheon Company "Fathometer Junior," at 5-second intervals; a stop watch was used for timing the intervals. At this sounding interval and ship speed the soundings were approximately 60 feet apart. Distance from shore directly astern was read off the radar at 0.1-mile intervals by the navigator. A student assistant recorded the soundings and distances as they were read off for later construction of profiles.

The total of 280 statute miles of profiles represents about 35 hours of sounding. Thus, the 106 profiles are based on an average of about 235 soundings each. Use of the ship's large recording echo sounder would have eased the field operation, but the reading and replotting of the records to take into account corrections for ship speed, tide height, and ship draft would have introduced more chances of error than did use of the small visual echo sounder. Checks of the echo sounder provided by wire depths obtained at a bottom-sample station between most adjacent profiles show that the sonic soundings are probably accurate to within about 3 feet.

Some of the sounding profiles were carried all the way into shore by taking lead-line soundings from a skiff in the shallow water; however, an unreasonably large amount of time would have been required for the ship to stand by for such soundings in all areas. For most profiles the inshore part was filled in by the scanty soundings given on U. S. Coast and Geodetic Survey smooth sheets. Profiles were continued onto the adjoining land to a height of 300 feet or for a distance of several hundred yards by using contours on the newly published quadrangles of the U. S. Geological Survey. On Santa Barbara Island, for which there is no good map, land profiles were measured with an altimeter. Although the profiles of the land are the best obtainable from existing sources, they are far less accurate than the profiles of the sea floor.

Presentation

Profiles connecting the soundings (Figs. 3–12) were drawn with an 8.0 vertical exaggeration, chosen as best able to reveal details without too great distortion. Even casual inspection of the profiles reveals the presence of flat portions, many of which would be acceptable to most geologists as terraces—for example on profiles 2, 3, 9, 61, 84–85, and 106. Profiles of Osborne, Fortymile, Cortes, and Tanner Banks clearly show flat tablelike terraces, and atop and near the middle of each bank is a steep nipple-like hill, evidently an erosional remnant above the shallowest terrace. Other profiles contain terraces that are less certain because of (1) gradual changes of slope (mostly near mouths of large rivers of the mainland—profiles 5 and 21, for example); (2) shortness of the flat portions (generally on steep shelves bordering islands—profiles 18 and 44, for example); (3) general irregularity of the whole profile (mostly on exposed sides of islands—profiles 35 and 55, for example).

Identification of terraces on the profiles is a subjective process, particularly for those described as less certain types. Therefore, in order to make clear to the reader which parts of the profiles were considered terraces, brackets (⌜ and ⌝) were drawn atop each profile to mark the landward and seaward ends, or shallow and deep ends. If the flattenings that were marked are true terraces and not haphazard decreases of slope, they should occur at similar depths on adjacent profiles; however, it is difficult to visualize correlations from profile to profile. What is needed is a memory-storing device similar to the one used for visualizing and comparing a series of soundings—in that instance a line connecting the soundings and called a profile. Such a device for comparing a series of profiles is illustrated in Plate 1.

The range of depth below sea level and height above sea level that is included within each

profile is represented by the narrow vertical lines of Plate 1. The depth, or height, span of each flattening of a profile or terrace is shown by broad lines extending between the landward, or top, end to the seaward, or bottom, end. Where both the top and the bottom ends of the flat-bank tops, reveal terraces not represented on near-by profiles; these have been allowed to pinch out in the simplest possible manner by appropriate placement of connecting lines. Correlation lines were omitted between profiles of distant islands.

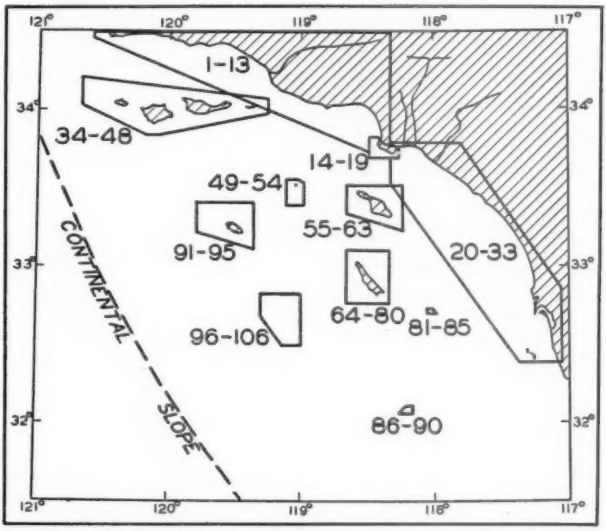

FIGURE 2.—INDEX MAP SHOWING SHAPES AND POSITIONS OF DETAILED MAPS
Numbers indicate the profile numbers included within each detailed map.

tening (edges of terrace) were considered definite, the span of the terrace is shown as a wide bar. Where the edges were considered to be less certain, the span is shown by a medium-width bar. Where either the top or bottom edge is definite and the other less certain, the span is indicated by a tapering bar. Terraces that have no obvious or recognizable top or bottom are shown by a triangle, the point of which faces the unknown edge. Terraces having no recognizable top are characteristic of the seaward side of San Clemente Island, where the shelf edge is a definite bottom edge of a terrace, but there is no concavity between this point and the bottom edge of the next-shallower terrace. Other examples exist where profiles were run into depths shallow enough to cross the bottom edge of a terrace but not shallow enough to reach the landward edge of the same terrace.

Possible correlations between terraces of adjacent profiles are shown by connecting lines, with a wide connecting line representing the shelf edge (shelf-break of Dietz and Menard, 1951). Several profiles, particularly those on

RESULTS

Depth Distribution

Plate 1 shows several sets of terraces that extend throughout the continental borderland. The shallowest terrace, which has an outer or deeper edge that averages about 40 feet, was reached by only 22 profiles, because of navigational danger to the ship; however, where soundings were carried inshore by skiff, as at the Palos Verdes Hills, it was found to be almost invariably present. The inner part of this terrace may be undergoing erosion at present, but the outer part is probably below effective wave base, too deep to cause wave refraction and much too deep to cause breaking, which is the chief point of transfer of energy to the bottom (Dietz and Menard, 1951).

The next deeper, or second, terrace has an outer edge that averages about 80 feet in depth along the mainland and the northern islands that are close to the mainland shore but averages 125 feet on banks and offshore islands. It certainly is too deep to be much affected by

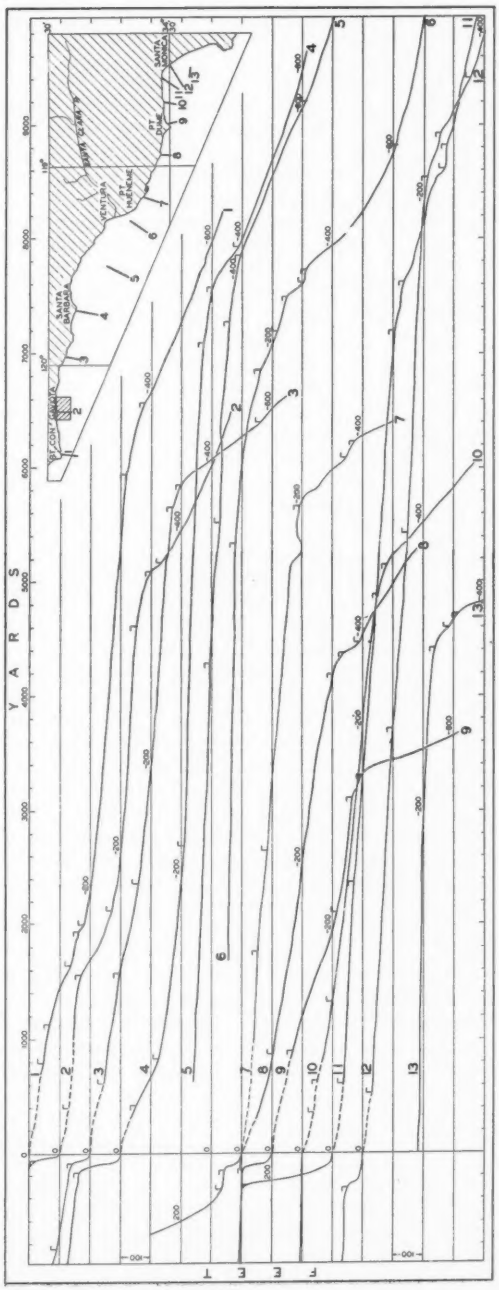

FIGURE 3.—PROFILES 1–13, POINT CONCEPTION TO SANTA MONICA

Each profile is a line connecting individual soundings at about 20-yard spacing. Except for profile 13, the shore line for each is indicated by the vertical line at the zero position of the distance scale. The extension of the profile onto land, shown to the left of the vertical line, was obtained from existing contour maps. The dashed line in shallow water is drawn on the basis of scanty chart soundings in depths too shallow to be reached safely by the ship. Vertical exaggeration is 8.0. Symbols ⌜ and ⌝ bracket portions of profiles believed to represent terraces. Cross-hatching at profile 2 shows area of contour map (Fig. 1).

wave erosion. This terrace is widespread throughout the region and can be identified on 48 of the 74 profiles that extended into shallow enough depths to reach it. It is absent on deep banks.

The third terrace is still more widespread and can be identified on 62 of the 91 profiles that reached its depth. Like the second terrace, the third one has an outer edge that is shallower along the mainland (160 feet) than around the offshore islands and banks (245 feet).

TABLE 1.—DEPTHS OF OUTER EDGES OF SUBMERGED TERRACES ON THE SHELVES OF SOUTHERN CALIFORNIA

(In feet)

Terrace	Mainland and nearshore islands	Offshore islands and banks	Difference
1	30?	50?	20?
2	80	125	45
3	160	245	85
4	255	345	90
5	290	430	140

Most common of all is the fourth terrace. It was identified on all except 3 of the 106 profiles, and on most profiles it is the widest of all the terraces. The depth of the outer edge averages about 255 feet along the mainland and 345 feet around the offshore islands and banks. In some places this terrace forms the shelf edge, but in most places the shelf edge is a compound feature which has a gentle break above a sharper one. The sharper break was taken as the outer edge of the fifth and deepest of the shelf terraces. It was recognized on 65 profiles at an average depth of 290 feet along the mainland and 430 feet on the offshore islands and banks.

In addition to the five main terraces, some smaller and more erratic ones are recorded by the profiles, particularly by profiles off the islands and banks, where masking by sediment from shore is less intense. These are shown as holes in the terrace patterns or as local terraces between the five main ones. Possibly, these erratic terraces represent brief erosion into especially soft rocks of only local exposure, or some may be slump blocks. The basin slopes beyond the shelf edge also contain some broad concavities that are similar to the terraces of the shelves except that they are much steeper. The steepness, irregularity in depth, and lack of an outer edge seems to preclude their being terrace remnants, and no connecting lines were drawn between them.

Warping

The depths of the terraces off southern California depend to a large extent upon their location in the region (Table 1). Along most of the mainland and the northern near-shore islands (Santa Rosa, Santa Cruz, and Anacapa) the terraces are shallower than off other islands and banks that are farther from the mainland shore. The difference in depth implies a relative downward warp in a seaward direction (or upward in a landward direction) with a greater warp for the deeper than for the shallower terraces. The ratio of the differences in depth to the offshore depths exhibits only a small range, 26 to 40 per cent. This difference cannot be the result of a deeper wave base offshore than nearshore, because wave base is only about 30 feet according to Dietz and Menard (1951); this is small compared with the depth differences being considered. Moreover, the depths on windward and leeward sides of most islands are similar. San Clemente Island may be an exception, but the depths on opposite sides of it may differ because of the local tilting suggested by Lawson (1893) and by Smith (1900).

The shelf edge, an easily identifiable point on each profile, serves as the best indicator of the nature of the warp. Average depths were determined for the shelf edge around each island, on each bank, and for groups of about five adjacent profiles of the mainland. These average values plotted in map form (Fig. 13) were contoured at a 25-foot contour interval; the resulting map of the depth distribution of the shelf edge clearly shows the nature of the warp. Supposedly unwarped areas may be present only north and south of the continental borderland off southern California. Although precise data on the depth of the shelf edge exist in neither area, the scanty information available indicates that the shelf edge in both area is about 400 feet deep, about the same as shown at the north and south limits of Figure 13; this suggests that the areas of shallower shelf edge have been warped upward. The terraces are of Pleistocene age and probably were cut during one or more lowerings of sea level during that epoch. According to Flint (1947, p. 437) the volume of ice at the maximum extent of Pleistocene glaciation produced a maximum lowering of sea level of about 400 feet, slightly less than Shepard's (1948, p. 143) figure of 430 feet for the average depth of the outer edge of the con-

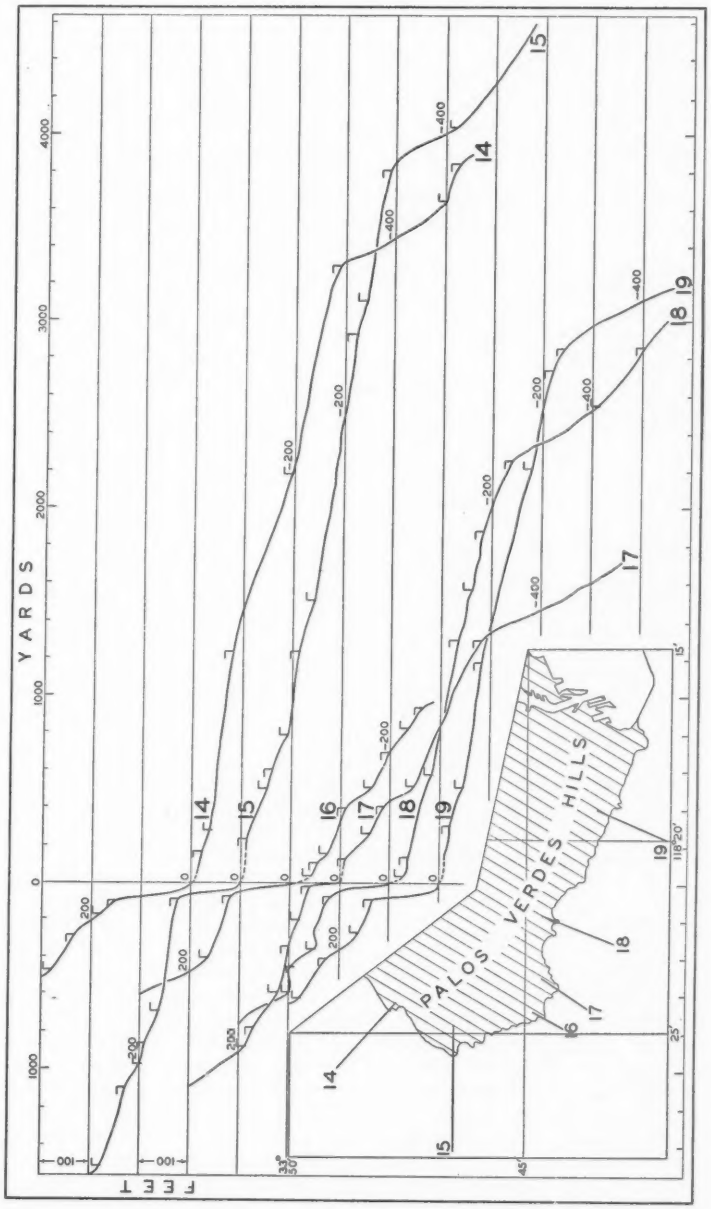

FIGURE 4.—PROFILES 14-19, PALOS VERDES HILLS
Symbols as in Figure 3.

tinental shelf around the world. If the shelf edge marks the outer edge of the erosional surface formed during the time of maximum lowering of sea level, then post-terrace warping must have been upward throughout most of the continental borderland in order to bring the shelf edge up to its present depth of only about 250 feet near Los Angeles. Downwarping, however, could have occurred south and west of the 400-foot contour (Fig. 13), which may thus indicate the approximate position of the hinge line of warping.

The average warp of the shelf edge, excluding the steepening near the continental slope, is about 160 feet per 100 miles. This is in the same general direction as the warp previously determined for the continental borderland on the basis of depths of hill tops, basin floors, and basin sills (Emery, 1954). The latter warp may have begun at about the same time as block faulting commenced, probably during Late Miocene, and long before the shelves were cut. It is about 3000 feet per 100 miles, about 20 times the amount of warp computed for the shelf edge. The difference is commensurate with the difference in age of the borderland topography and of the shelves, and it suggests that warping has gone on more or less continuously since the borderland was formed up to the present. This warping of the continental borderland may be related to the arching of the Transverse Ranges shown by Gilluly (1949) to amount to about 20 inches per century. In addition to the general regional warping, local folding and faulting has occurred—as at Ventura (Putnam, 1942; Bailey and Jahns, 1954)—but such folding and faulting evidently has been so local that it has not affected the general picture.

Comparison with Other Regions

The first, or shallowest, terrace has its counterpart in many other places of the world. In the Pacific Ocean it has been reported from Oahu in the Hawaiian Islands (Stearns, 1935); Espiritu Santu in the New Herbrides (Stearns, 1945); Bikini, Eniwetok, and Rongelap atolls in the Marshall Islands (Emery, Tracey, and Ladd, 1954); Raroia atoll in the Tuamotu Islands (Newell, 1956); Japan (Komukai, 1956); and Guam in the Mariana Islands (Emery, in press). In the Atlantic Ocean it has been reported off Trinidad (van Andel and Postma, 1954, p. 19); from Bimini and Andros Islands in the Bahamas (Newell and Imbrie, 1955); on banks off the coast of Texas (Parker and Curray, 1956); and Bermuda (Heinz Lowenstam, personal communication). In the Indian Ocean it has been mapped in the Persian Gulf (Houbolt, 1957). Probably it has been noted in many other areas but not reported. Certainly it is a world-wide feature.

Deeper terraces on the continental shelf are far less well known than the one cited above, owing to greater difficulty of obtaining precise data. One of the first and outstandingly good reports of deeper terraces was given by Yabe and Tayama (1934) who reported from studies of navigational charts around Japan and Korea the presence of nine submarine terraces with outer edges at 15 feet (only in Yellow Sea), 65, 100, 195, 330, 460, 750, 1150, and 2620 feet. The three deepest ones are beyond the shelf edge and are less widely distributed than the others. Three years later and quite independently Shepard and Wrath (1937) examined many navigational charts of all oceans and noted that the mean depths of benches or flat areas of the shelf are most common at depths of about 45, 95, 155, 205, 260, and 300 feet, in contrast to the average depth of the shelf edge at 430 feet (Shepard, 1948, p. 143). Veatch and Smith (1939, Pl. IB, IIB, IIIB) named a Franklin Shore that rises from a depth of about 360 feet off Long Island to about 270 feet north of Chesapeake Bay. They also indicated a Nicholls Shore at about 480 feet off Long Island. Cooper (1948) reported a submerged sea cliff off England, which has a base at 145 feet and is fronted by a broad terrace. Tayama (1950), continuing his earlier work, described in detail the sea floor southwest of Tokyo and recorded submarine terraces on the shelf with outer edges at 35, 80, 195, and 345 feet and one terrace seaward of the shelf edge with its edge 690 feet deep. Studies of the continental shelf off Western Australia by Carrigy and Fairbridge (1954) show terraces which have outer edges at 30, 70, 150, 210, 330, 360, 450, 510, 570, and 600 feet. Parker and Curray (1956) reported that the least depths of 130 banks that rise above the continental shelf off Texas are concentrated at about 55, 200, and 280 feet. In the Persian Gulf Houbolt (1957) identified terraces at 55, 100, 170, and 240 feet. Komukai (1956) mapped in great detail terraces at 65, 140, 230, 280, 310, 360, and 460 feet between the Japanese Islands of Honshu and Hokkaido. Finally, in 1952 at Guam Emery (In press) discovered terraces having outer edges at depths of 55, 105, 195, and 315 feet.

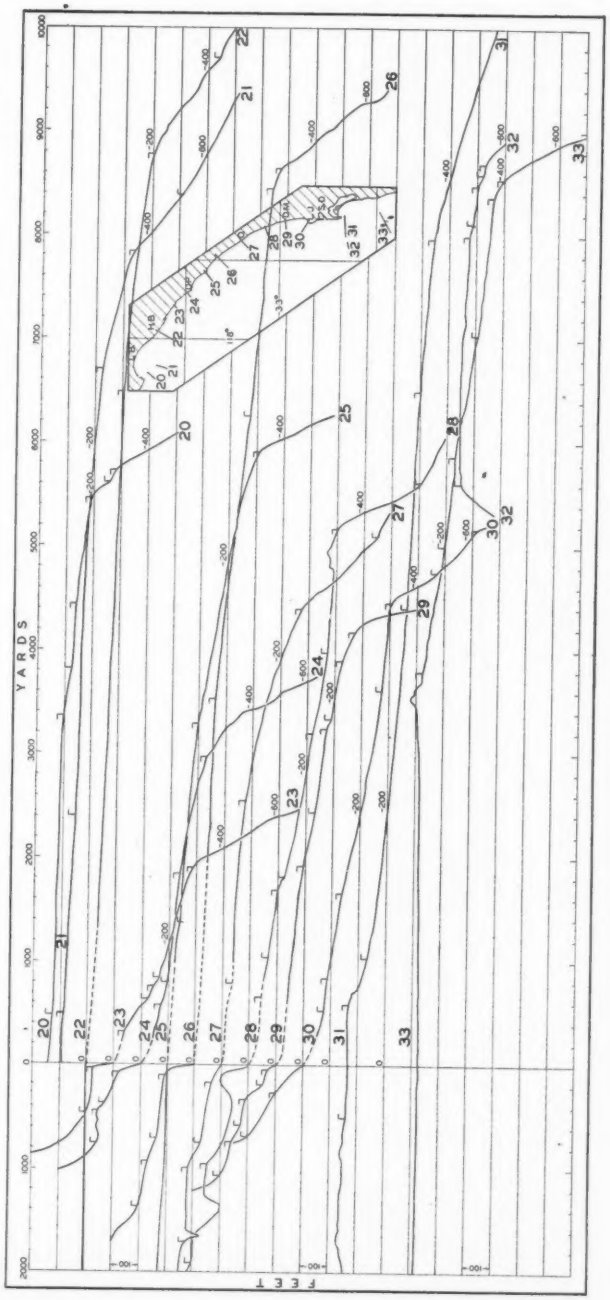

FIGURE 5.—PROFILES 20–33, LONG BEACH TO LOS CORONADOS ISLANDS, MEXICO
Symbols as in Figure 3.

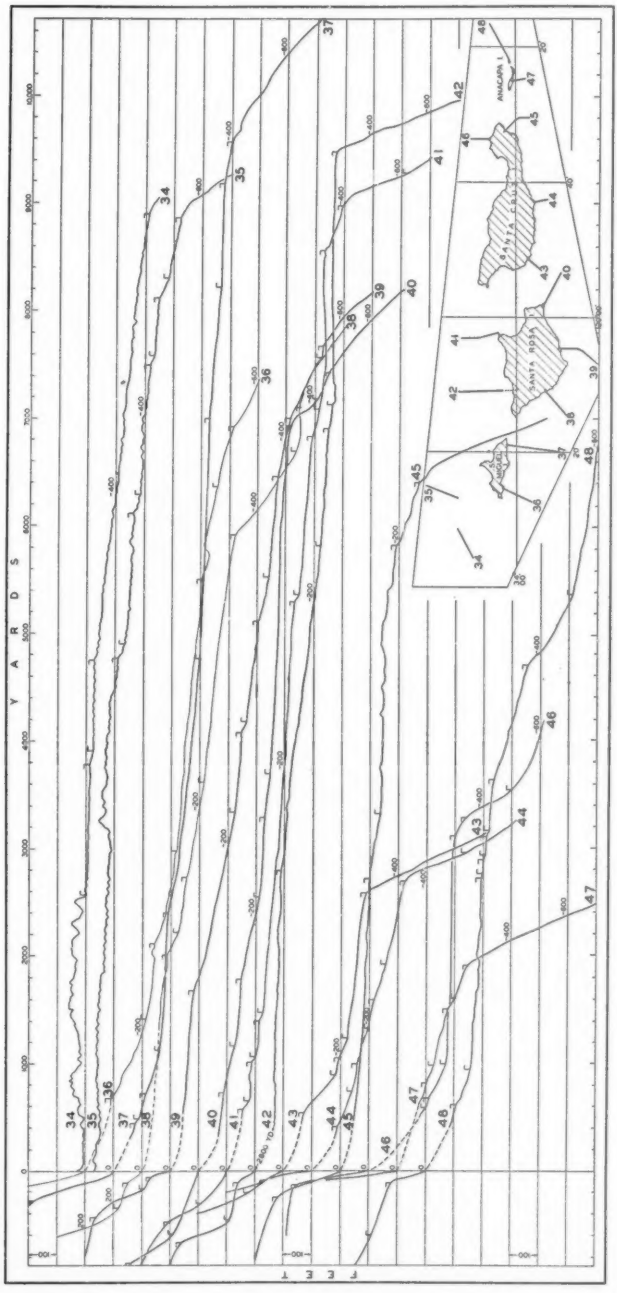

FIGURE 6.—PROFILES 34–48, SAN MIGUEL TO ANACAPA ISLANDS
Symbols as in Figure 3.

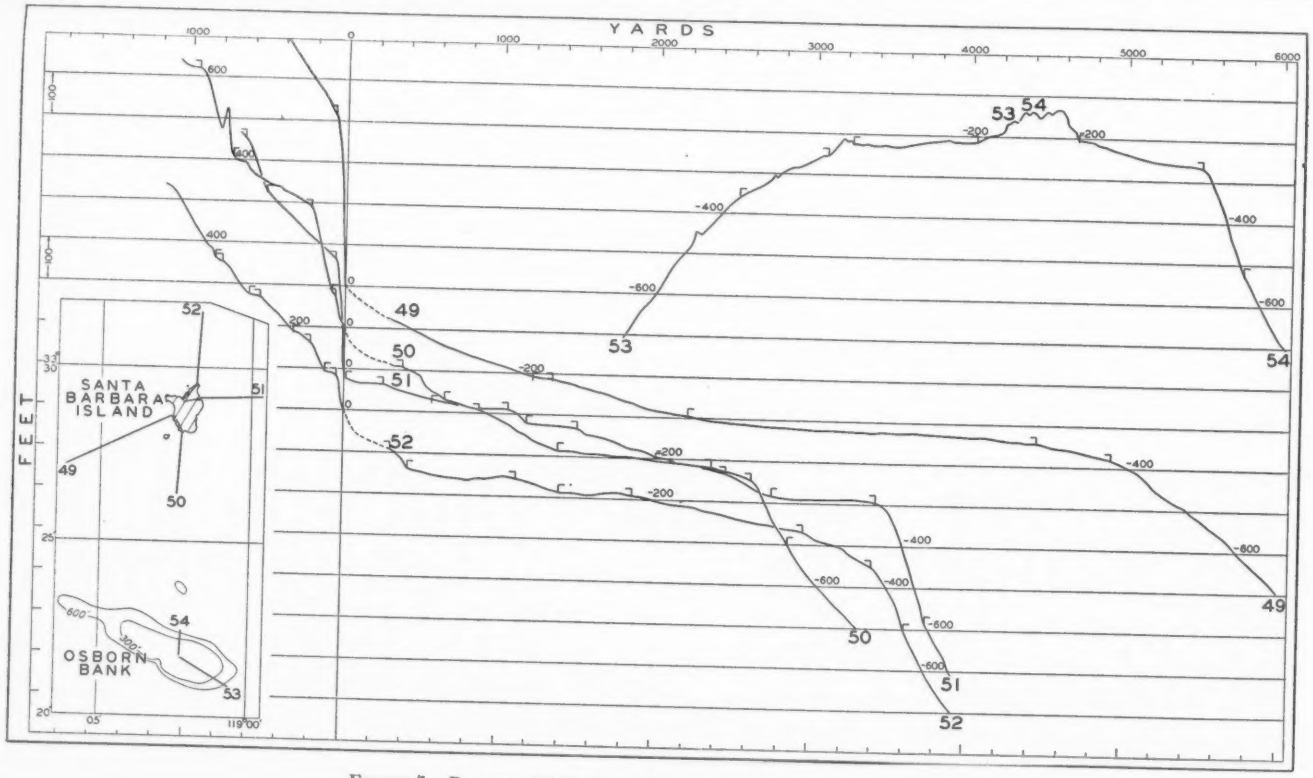

FIGURE 7.—PROFILES 49–54, SANTA BARBARA ISLAND AND OSBORN BANK
Symbols as in Figure 3.

Reported depths of the outer edges of submerged terraces suggest that they may be grouped into the following most common depth ranges: 45–65, 80–125, 170–205, 230–280, 300–350, and 430–460. Little confidence, however, should be placed upon such widespread and general correlation, particularly since deformation has altered their depths in some areas, such as southern California. Yet, it is significant that where detailed examinations of the continental shelf have been made, several terraces can be identified. This points strongly toward an origin related to eustatic changes of sea level, even though the number and vertical range of eustatic changes cannot be considered as established.

Composition

At a lowered sea level a terrace could be formed either by erosion at the new wave base or by deposition through deltaic growth. In southern California there is abundant evidence that the terraces are underlain by rock. This is known in many areas from dredging work, oil company jet-coring explorations, underwater photographs, and records obtained by a powerful echo sounder capable of obtaining reflections from strata beneath the sea bottom. In addition, divers report the observation in a few places of aligned rock outcrops at the outer edges of shallower terraces. These data do not exclude sediment-built platforms, because prograding of steep slopes can occur (Emery and Terry, 1956). However prograding is common probably off only a few parts of the mainland coast where the supply of detrital sediments is relatively large and is rare off islands and banks.

The chief role of sediments is believed to be that of burying and obscuring the terraces rather than forming them, as implied by the longer and gentler slopes and the lack of sharpness of terraces in areas off large rivers such as the Santa Clara, Los Angeles, San Gabriel, and Santa Ana (indicated from left to right in Figure 2). In contrast, the profiles are sharp and contain several small extra terraces off islands and banks where the supply of sediment is very small. Additional evidence of burial is provided by studies of the sediment thickness on the shelf by W. C. Thompson (Dietz and Menard, 1951), who showed the presence of a filet of sediment occupying the notch of a terrace off Santa Barbara.

Rocks that have been truncated by the terraces are not merely flat-lying unconsolidated sediments. They consist of volcanic rocks (as at San Clemente Island), metamorphic rocks (as off part of Santa Catalina Island and Sixty-mile Bank), and folded Miocene and older sedimentary rocks at most other places (Emery and Shepard, 1945). The total width of the submerged terraces exceeds the total width of the terraces that have been raised above sea level in most localities, including the Palos Verdes Hills and San Clemente Island, and although the raised terraces obviously required a long time to be cut, the wider submerged ones needed a correspondingly longer time. Moreover, the depth differential between the same submerged terraces along the mainland and the offshore islands and banks shows that they probably were not cut quickly in a matter of only a few years but that cutting was spaced throughout the whole period of warping.

Date of Terrace Cutting

Locally, as at San Pedro Bay (Moore, 1954) just east of the Palos Verdes Hills, and also along the coast between Ventura and Santa Barbara, the submerged terraces were cut across Pliocene shales, which shows that they are post-Pliocene. Supporting evidence of a Pleistocene age for the terraces is provided by the presence of deeply filled stream channels along the whole coast. These channels contain as much as 230 feet of fill (Pl. 1) where they cross the shore line. Some of them, those that cross the Los Angeles and Ventura basins, are known to be post-Pliocene because they cut Pliocene sediments, and some by the same evidence are post-Early and even post-Middle Pleistocene (Upson, 1949; Crowell, 1952; Poland, Piper, and others, 1956). However, the great depth of most of the terraces below wave base shows that they were not formed at the present position of sea level.

Although the age of the terraces is evidently post-Pliocene and pre-Recent, the important question remains of whether they were cut during several pauses of a generally rising sea level during the last glacial age or whether they represent the lowest positions of sea level for each of the main Pleistocene glacial ages. Points which favor the first possibility are (1) the known instability during the mid-Pleistocene of parts of southern California, such as the Ventura region (Eaton, 1943; Putnam, 1942); (2) the post-mid-Pleistocene age of some of the stream valleys of the Los Angeles and Ventura basins that were incised to terrace levels and later filled with sediments; and (3) the possible multiple lowerings of sea level

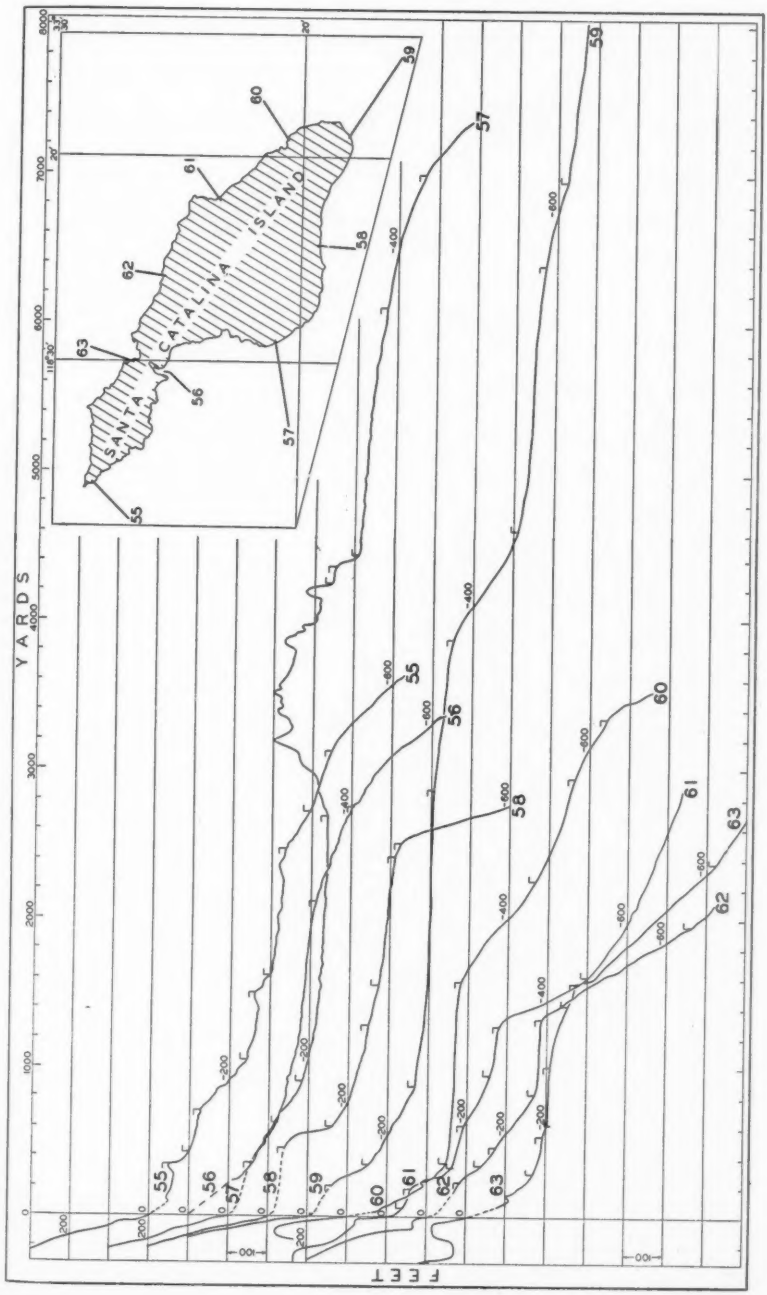

FIGURE 8.—PROFILES 55–63, SANTA CATALINA ISLAND

Symbols as in Figure 3. These profiles were plotted by Mr. E. C. Tripp as a term paper for the course in Marine Geology at University of Southern California.

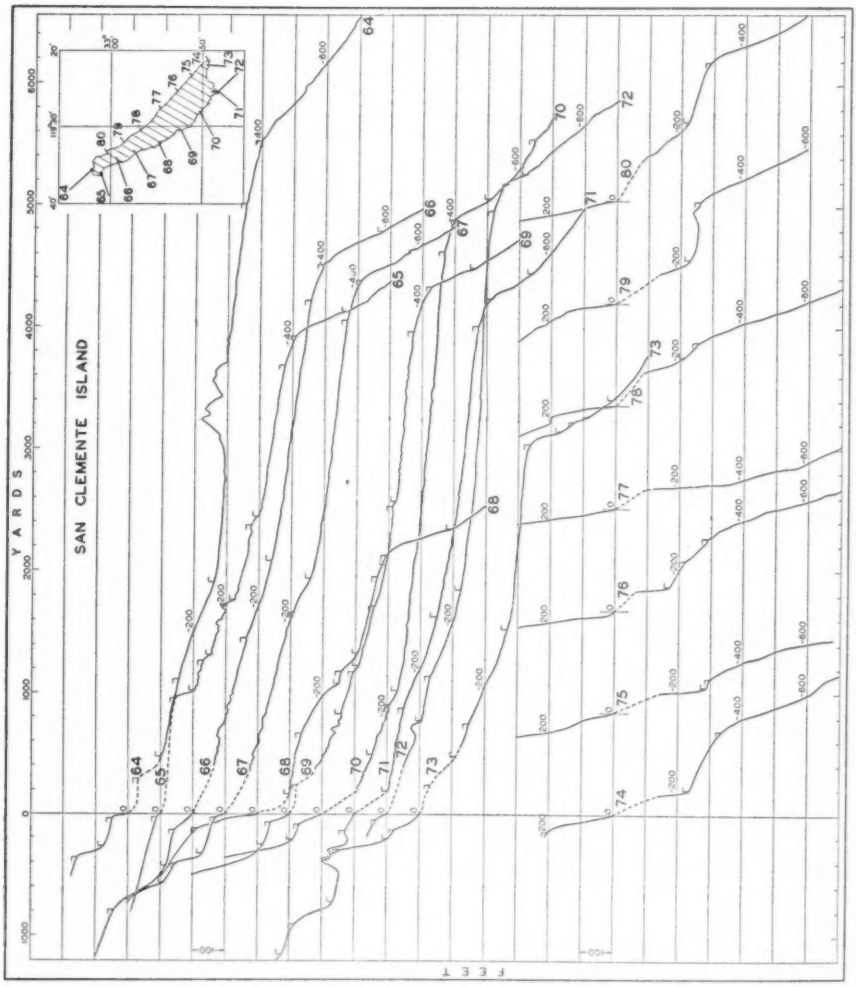

FIGURE 9. PROFILES 64–80, SAN CLEMENTE ISLAND
Symbols as in Figure 3.

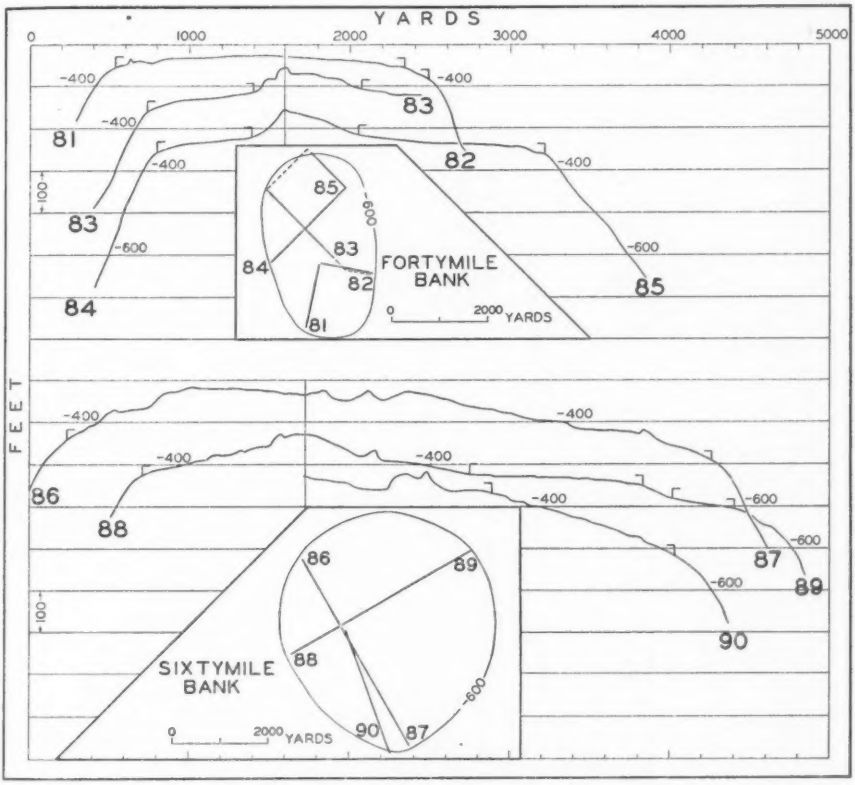

FIGURE 10.—PROFILES 81–90, FORTYMILE AND SIXTYMILE BANKS
Symbols as in Figure 3.

FIGURE 9. PROFILES 64–80, SAN CLEMENTE ISLAND
Symbols as in Figure 3.

during the last glacial age suggested by Zeuner (Ericson and Wollin, 1956) and indicated by the separate subages Mankato, Cary, Tazewell and Iowan (Flint, 1947, p. 210–212) based on distinct sheets of drift in the Illinois region, and by the separate moraines in the Sierra Nevada that may be synchronous with separate terracē levels of ancestral Mono Lake (Putnam, 1950). Points which favor cutting of the terraces throughout the Pleistocene are: (1) the average width of 4 miles and the maximum width of 15 miles of the shelf seem to require much longer than the possible 50,000 years of Wisconsin time, especially in view of the very narrow surface which is now being cut across rocks of equal resistance; (2) the progressively greater warp of deeper than of shallower terraces, implying the passage of considerable time between cutting of successive terraces.

The fact that there are five terraces is not very helpful because it corresponds closely to the number of glacial ages of the Pleistocene and the number of glacial subages of the Wisconsin. Similarly, the presence of similar groupings of terraces throughout the world merely indicates eustatic origin and cannot distinguish between Pleistocene and Wisconsin duration until Pleistocene diastrophism has been more accurately dated.

If the shelf is a feature formed by many lowerings of sea level, during each of which additional rock was planed off to form the present wide expanse, then it must be considered very doubtful that a terrace and sea cliff formed at a given maximum lowering would remain uneroded by the next lowering of sea level. Thus, it seems unreasonable that each terrace marks a eustatic level of a separate glacial age. The writer, therefore, favors the concept that the

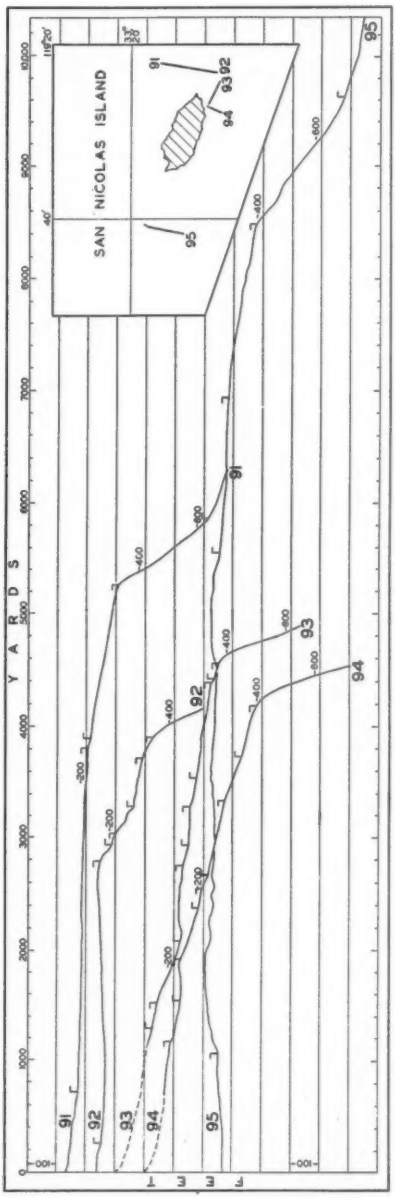

FIGURE 11.—PROFILES 91–95, SAN NICOLAS ISLAND
Symbols as in Figure 3.

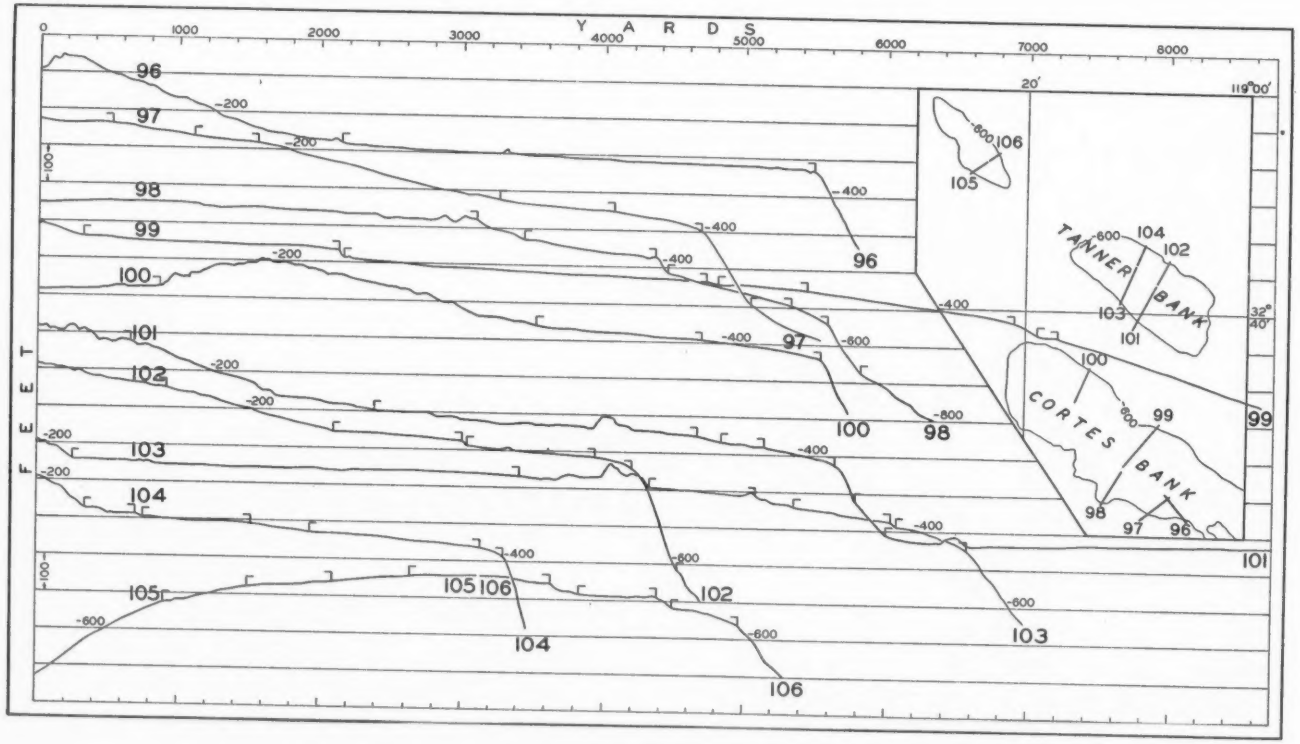

FIGURE 12.—PROFILES 96–106, CORTES, TANNER, AND A NEAR-BY UNNAMED BANK
Symbols as in Figure 3.

shelf width was produced by erosion during several glacial ages but that the terraces are features eroded into the shelf by successively lesser eustatic lowerings of sea level during the Wisconsin subage.

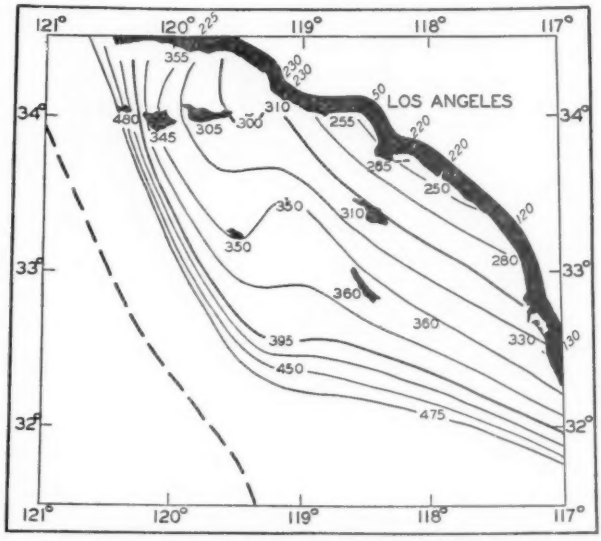

FIGURE 13.—DEPTH DISTRIBUTION OF SHELF EDGE SHOWN BY 25-FOOT CONTOUR INTERVAL
Vertical numbers indicate average local depth of shelf edge below sea level obtained by averaging shelf edge of 5 or more adjacent profiles. Slant numbers show depths to which stream valleys have been cut below sea level and filled with sediments (obtained from various sources).

In a study of the sediments of the basin slope off the Palos Verdes Hills Emery and Terry (1956) obtained a core which had near its bottom a layer of pebbles encrusted with dead calcareous red algae. Subsequently, the algae were found in four other cores, all within 0.4 miles of each other and between depths of 334 and 387 feet near the end of profile 19. Because these depths are too great for growth of algae in the somewhat turbid water of the area, it is believed that the algae lived when sea level was at or near the depth at which the lowest terrace was cut. The depth then would have been about 130 to 180 feet, shallow enough to permit growth of the algae. The possibility of distant transportation of the algae is ruled out by their fragile nature and by the fact that many of the pebbles are completely encrusted. At some of the core sites the algae were subsequently buried under as much as 120 inches of detrital silts, clays, and sands. Radiocarbon dates show a range from 17,000 to 24,500 years (Table 2). This supports the view that the terraces are Wisconsin age rather than extending back to Early Pleistocene. An element of uncertainty results from the possibility that the algae could have been deposited during a time of lower sea level too brief to have permitted the cutting of a terrace; in addition, the precise depth of water at the time of algal growth is unknown. If the radiocarbon measurements are accepted as indicating the date of cutting of the deepest shelf terrace, then this terrace is not only Wisconsin but is younger than the +75-foot terrace on shore of the Palos Verdes Hills, which was shown by Kulp, Tryon, Eckelman, and Snell (1952) to be overlain by shells having a radiocarbon age greater than 30,000 years.

Relationship to Very High and Very Deep Terraces

Terraces (poorly preserved in most places) that stand 1000 feet or more above sea level occur at Palos Verdes Hills (Woodring, Bramlette, and Kew, 1946), San Clemente Island (Lawson, 1893; Smith, 1900), Ventura (Putnam, 1942), and many other places of the mainland and islands. Because this elevation is far too great for them to be of eustatic origin,

they must have risen to their present positions by diastrophic movements. Similarly, on the sea floor are other very deep terraces (Shepard and Emery, 1941) and accumulations of rounded gravels far below sea level (Clements and Dana, 1944; Emery and Shepard, 1945) that also must have reached their present depths by diastrophism. Most of these very high and very deep terraces cannot be older than Middle or Late Miocene, because the terraces are cut across and their gravels consist of rocks that are of that age. The much greater erosion of the high terraces, the very high and very deep terraces above and below accepted Pleistocene eustatic sea levels, and the greater difficulty of correlation of the very high and very deep terraces, indicate that these terraces must be older than the ones atop the shelves. If the shelf terraces were formed by eustatic levels related to all glacial ages of the Pleistocene, or if the shelf were beveled by each glacial lowering of the Pleistocene, then the very high and very deep terraces must be pre-glacial, or Pliocene, a suggestion already made by Upson (1951) and others. If the shelf terraces are Wisconsin the very high and very deep terraces could be as young as Middle or Early Pleistocene; this conclusion is supported by much paleontological and structural data at Ventura and Palos Verdes Hills, although geological data in most other areas would permit them to be older. Such a possible wide variation in age of the high terraces might help to solve the problem of why high terraces are well preserved on Palos Verdes Hills and San Clemente Island and poorly preserved on the intervening Santa Catalina Island (Smith, 1933), though all three areas have excellent low terraces on land and others on the shelves.

Future Studies

Perhaps when more is learned of the regional picture of raised and submerged terraces a better understanding of eustatic sea-level changes of the Pleistocene can be reached. At present the regional study of the raised terraces has lagged behind their detailed study in local areas. Information about the structures truncated by the raised terraces, the nature of the post-terrace fill, and the age of enclosed fossils is obtainable far more easily than for the submerged terraces, as shown by studies of the terraces on the Palos Verdes Hills by Woodring, Bramlette, and Kew (1946). More such studies and studies on a regional basis are needed. On the other hand, the submerged terraces may provide more reliable information regarding eustatic changes of sea level because these terraces have probably undergone less postcutting erosion than those on land. In any event it seems obvious that glacial and interglacial terraces are so closely related that they must be studied jointly. Local departures from uniformity of the depth and height distribution of the terraces should then reveal much about the extent and date of Pleistocene deformation in coastal areas. Perhaps solution of the problem of history and time of terrace cutting must await the development of some form of isotope dating capable of satisfactorily spanning the required time interval.

Table 2.—Radiocarbon Dates on Calcareous Algae in Cores

Determinations by Magnolia Petroleum Company

Sample	Composite of*		Age
	Core Number**	Depth *in situ* inches†	
1	2981	118–128	20,000 ± 1000
	5158	58–60	
2	5156	98	24,500 ± 1600
	5157	56, 70	
3	5156	0–4, 34, 42, 60–62	18,600 ± 1200
4	5154	0–20	17,000 ± 1400

* Compositing was necessary to obtain sufficiently large samples for analysis. Samples 1 and 2 consisted of completely encrusted pebbles; samples 3 and 4 of broken and slightly reworked material.

† *In situ* depth is taken as 2.0 times depth in the cores obtained by simple gravity order.

** Core 2981—Lat. 33°41.5′ N.; Long. 118°19.9′ W.; 334 feet.

Core 5154—Lat. 33°41.6′ N.; Long. 118°19.7′ W.; 380 feet.

Core 5156—Lat. 33°41.6′ N.; Long. 118°19.9′ W.; 354 feet.

Core 5157—Lat. 33°41.6′ N.; Long. 118°20.0′ W.; 387 feet.

Core 5158—Lat. 33°41.3′ N.; Long. 118°19.6′ W.; 345 feet.

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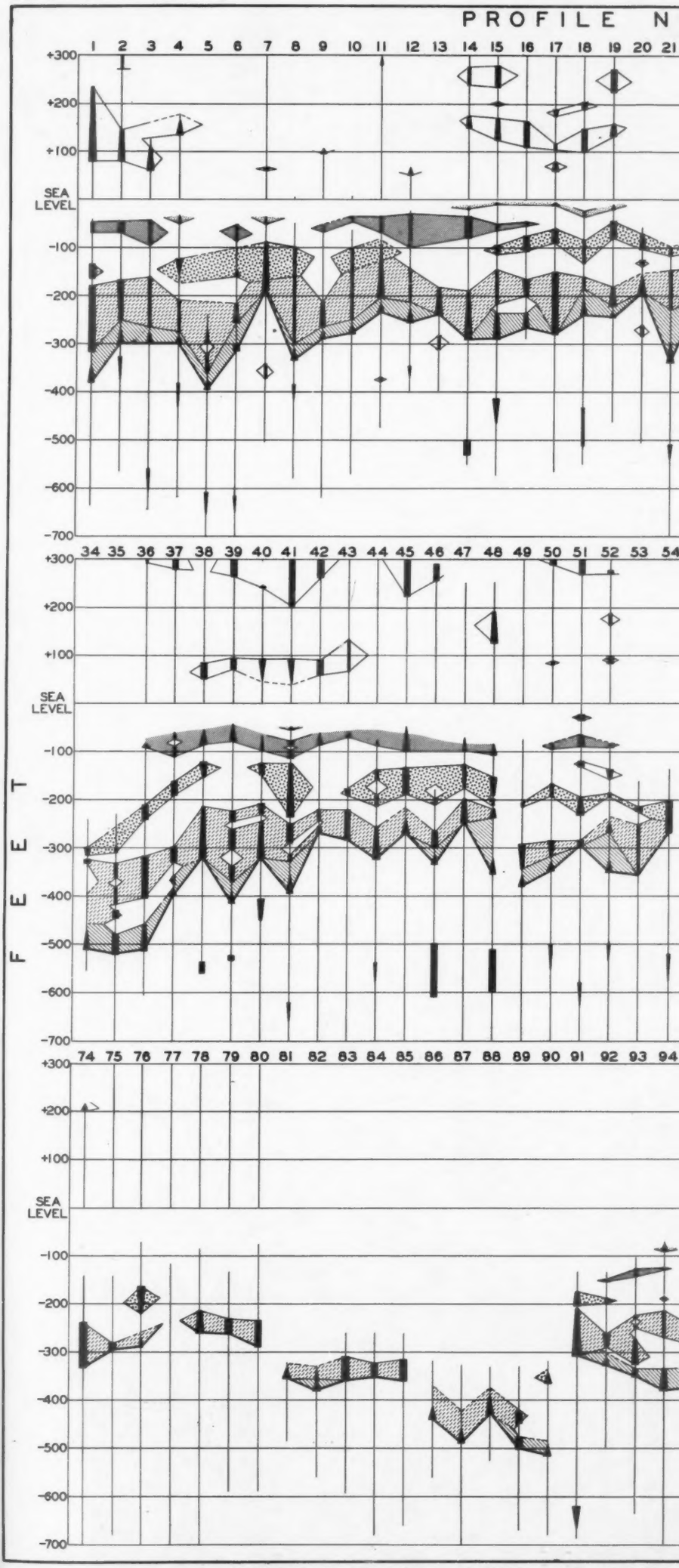
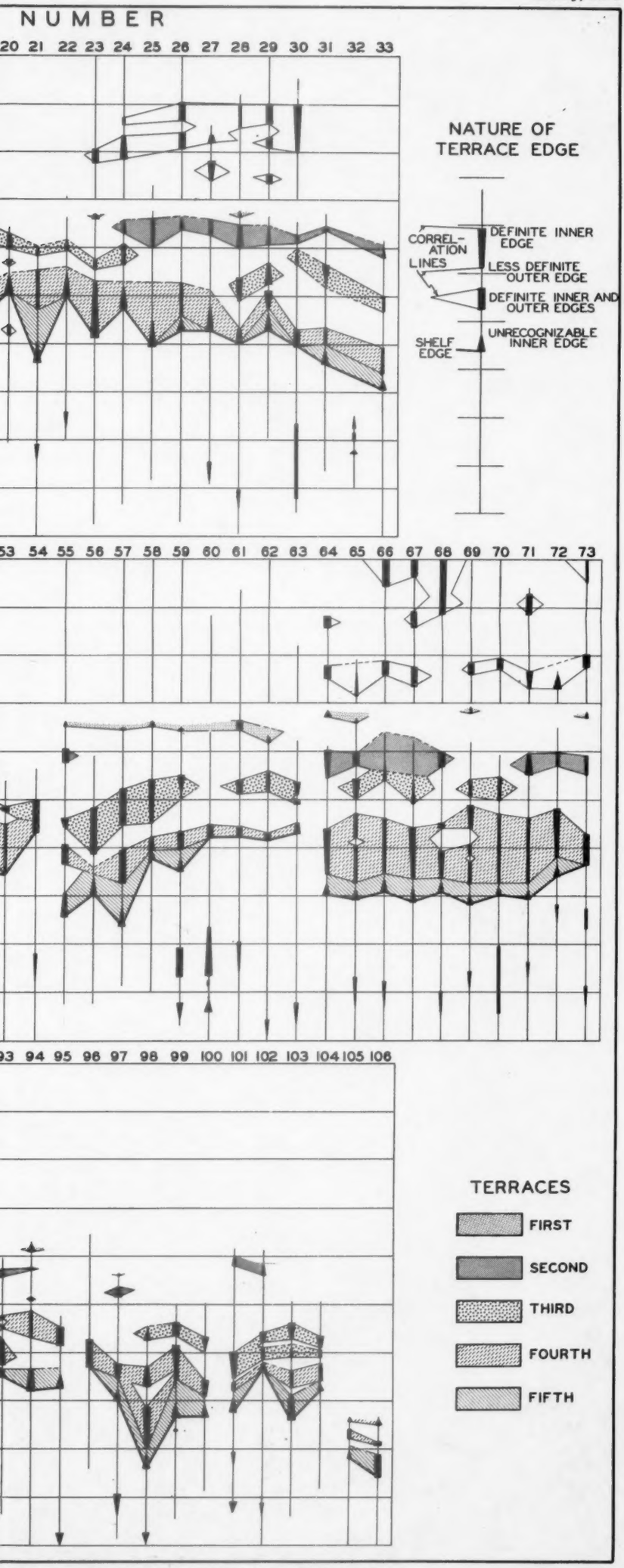


SUMMARY CHART OF TERRACE DEPT

Narrow vertical lines show depths below and heights above sea level as indicated by profiles of Figures 3-12. Wider portio terrac


NUMBER
20 21 22 23 24 25 26 27 28 29 30 31 32 33
53 54 55 56 57 58 59 60 61 62 63 64 65 66 67 68 69 70 71 72 73
93 94 95 96 97 98 99 100 101 102 103 104 105 106
NATURE OF TERRACE EDGE
CORREL-ATION LINES
DEFINITE INNER EDGE
LESS DEFINITE OUTER EDGE
DEFINITE INNER AND OUTER EDGES
SHELF EDGE
UNRECOGNIZABLE INNER EDGE
TERRACES
FIRST
SECOND
THIRD
FOURTH
FIFTH

EPTHS SHOWN ON ALL PROFILES

portions of vertical lines show depth or height range of terraces, based on symbols [and] used on Figures 3-12.

Introduc
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INTRO

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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ALTERATION OF CLAY MINERALS IN ILLINOIAN TILL BY WEATHERING

BY JOHN B. DROSTE AND JAMES C. THARIN

ABSTRACT

Alteration of clay minerals by weathering in Illinoian till of the Northwest Allegheny Plateau was determined by X-ray techniques. Analysis of closely spaced samples throughout the weathering profile from fresh till to the A-soil zone shows that each zone of the weathering profile is characterized by a specific clay mineral assemblage. Chlorite in the fresh till is degraded by hydration of the brucite sheets which leads to the development of vermiculite in the upper soil zone. Well-crystallized illite of the fresh till is degraded by hydration of the intermica sheets, and a random mixed layering of illite and montmorillonite is produced.

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INTRODUCTION AND ACKNOWLEDGMENTS

This paper analyzes the clay mineral alteration produced by the weathering of an Illinoian till near Titusville in northwestern Pennsylvania and compares it to the alteration of clay minerals produced by the weathering of Wisconsin tills in adjacent areas. The profile described is the only complete weathering profile developed on Illinoian till that has been discovered in northwestern Pennsylvania during the summers of 1954–1956.

The glacial deposits in northwestern Pennsylvania were studied between 1850 and 1902 (Leverett, 1934, p. 1–10). Leverett (1902, p. 21) had designated the deposits in the area of the section here as Kansan or pre-Kansan, but in 1934 he (p. 11) concluded that they were Illinoian and pre-Illinoian. The writers agree with this designation.

This work is part of a larger study of northwestern Pennsylvania Pleistocene geology by Prof. G. W. White and his associates and was supported by a National Science Foundation grant to Professor White. The authors express their thanks for this financial assistance in the field and to Prof. Ralph E. Grim and Dr. William F. Bradley for counsel in the laboratory study.

SECTION DESCRIPTION AND SAMPLING

Samples were taken from the zones into which tills of Illinoian age may be divided (Leighton and MacClintock, 1930; White, 1939, p. 164). The outcrop occurs in Cherrytree Township, Venango County, along Pennsylvania Highway 8, about 1.6 miles south of the Post Office in Titusville, Pennsylvania, and 1.2 miles southwest of the Drake Oil Well. Although loess or similar deposits are possibly present at the surface in the area of the out-

crop, the authors do not believe that any such deposits are present in the section sampled for this study.

A description of the section follows:

	feet	inches
Zone I		
(1) Loam, silty, gray brown; A-1 soil (not sampled because material disturbed by plowing).	0	9
(2) Loam, silty, dark dusky yellow; A-2 soil (sample 328).	0	7
(3) Loam, silty, clayey, moderate orange and gray mottling, small weathered siltstone and shale pebbles; upper B-1 soil (sample 329).	0	6
(4) Loam, silty, clayey, pronounced orange and gray mottling, nusiform fracture, small weathered sandstone, siltstone, and shale fragments; middle B-1 soil (sample 330 at 3 feet from surface).	1	7
(5) Loam, silty, clayey, pronounced orange and gray mottling, compact with some prismatic structure, well-weathered crystalline rocks, sandstone, siltstone, and shale fragments (sample 331).	1	0
Zone II		
(1) Till, silty, clayey, thoroughly weathered, orange and gray stains in joints, manganese stain coating joint surfaces, weathered crystalline rocks, siltstone, and shales (sample 332 at 5 feet 9 inches from surface).	1	11
Zone III		
(1) Till, as below but not calcareous; (sample 333 at 7 feet 7 inches from surface; sample 334 in contact zone between this unit and subjacent unit).	2	3
Zone IV		
(1) Till, silty, moderately pebbly, calcareous, moderate yellow brown, rough horizontal fracture; till matrix is slightly calcareous, but many small carbonate pebbles react violently to acid. Coal fragments common in places (sample 335 taken at 12 feet 5 inches from surface; sample 336 taken in contact zone between this zone and subjacent zone).	7	4
Zone V		
(1) Till, as above but light olive gray; many joints are present, along which ground water has oxidized the gray till in zones varying in thickness from 1/16 of an inch to 1½ inches. (Sample 337 taken from solid-gray area at 20 feet from the surface).	8	6
Sand and gravel, brown, calcareous, uneven contact with superjacent unit.	7	6

Bottom of cut at 31 feet, 11 inches

The authors know of no other section in this area in which calcareous Illinoian till is exposed. This deposit was preserved because of its deposition in a rather narrow channel in the bedrock. Elsewhere the till is so thin that no calcareous (horizon IV or V) till can be found.

Laboratory Procedure

Oriented aggregates (Grim, 1934, p. 45) of the <2-micron fraction, which was separated from each sample by sedimentation in water without a dispersing agent and without acid leaching to remove carbonates, were mounted on glass microscope slides.

Spectrometer traces were made on a General Electric XRD-3 recording diffractometer using filtered copper radiation. In addition to air-dried slides, spectrometer traces were obtained from oriented slides after glycolation and after heating to 450°C. Traces were also obtained from oriented slides after NH_4Cl treatment, as suggested by Walker (1951, p. 203).

Differential thermal analyses were made on the <2-micron fraction of each sample. Single clay mineral species lend themselves well to differential thermal analysis, but the mixture of clay minerals in these samples, particularly the mixed layers, made it impossible to obtain significant information by differential thermal studies.

Sand-silt-clay ratios were determined for each sample following the procedure described by Shepps (1953, p. 36).

Clay Mineralogy

All clay minerals are identified more accurately and more easily by examination of their basal (001) diffraction lines. The identification of each clay mineral type and the discussion of clay mineral type presented here is based on diffraction data obtained from basal planes of each sample.

The following is a detailed description of the clay minerals in each sample, starting with the unaltered material; changes due to weathering are then pointed out. In the discussion following

the description of the samples a comparison is made between the clay mineral alteration in an Illinoian and Wisconsin weathering profile.

Sample 337: gray calcareous unaltered original till, Zone V (Fig. 1).

(1) 14 Å peak. A sharp moderately intense peak which is unaffected by heating to 450°C. and glycolation occurs at 14.2 Å. This d spacing with appropriate higher orders,is attributed to well-crystallized chlorite, probably rich in iron. Chlorite makes up approximately 40 per cent of the clay size fraction.

(2) 10 Å peak. The strong sharp peak at 10 Å does not change after heating or after glycolation. This peak and its appropriate higher orders indicate the presence of well-organized illite. In some samples it is difficult to differentiate illite from well-crystallized micas (Grim, 1953, p. 93), but in these samples the characteristics of the 10 Å mineral indicate the presence of illite rather than mica. Illite comprises about 60 per cent of the clay size fraction.

(3) 7 Å peak. The problem of identifying kaolinite in the presence of chlorite arises here. The 7 Å/14 Å intensity ratio in this sample is 1/.6. The intensity of higher orders of the 14 Å peak (not shown in Fig. 1) and the argument previously presented (Droste, 1956, p. 915) leads to the conclusion that kaolinite is not present in the sample, or the percentage of kaolinite, if present, is so small that it cannot be detected.

Sample 336: Transition zone between gray (Zone V) and brown (Zone IV) calcareous till.

(1) 14 Å peak. The 14 Å peak in the untreated slides in this sample is broader and less intense than in sample 337. The peak does not shift after glycolation and NH_4^+ treatment and is almost completely lost after heating to 450°C. This is the first evidence of alteration of clay minerals. The nature of the alteration is presented in the discussion of sample 335.

(2) 10 Å peak. The illite peak in this sample is as intense as it was in 337, but slight asymmetry is present on the low-angle side. This asymmetry is lost after heating to 450°C. and suggests that the illite is being degraded. This indicates, also, that clay minerals alter very early in soil development.

Sample 335: Calcareous brown till, Zone IV (Fig. 1).

(1) 14 Å peak. The peak in the untreated sample is broad and about as intense as that in sample 336. It is all but lost and is shifted to 13.3 Å after heating to 450°C., and the peak is changed to a plateau after glycolation. Its shape or position does not shift noticeably after NH_4^+ treatment. This is interpreted to mean that a few hydroxyls in the brucite sheets are joined by protons, but that the major brucite configuration is maintained. This degrading of chlorite, which began in the underlying zone, continues through successive zones to be described.

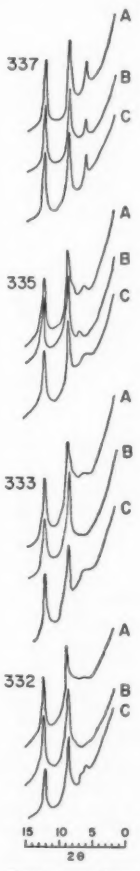

FIGURE 1.—SMOOTHED SPECTROMETER TRACES OF SAMPLES 337–332

Made on an XRD-3 General Electric recording spectrometer using filtered copper radiation. (A) sample air dried; (B) sample heated to 450°C; (C) sample treated with ethylene glycol.

(2) 10 Å peak. The increased asymmetry on the low-angle side of the 10 Å peak, and the loss of this asymmetry after glycolation and heating to 450°C. indicates further degrading of the illite, probably by removal of K^+ from between the sheets. This degrading began in the previous sample and continues in samples to be discussed. It is important to point out that degradation of both illite and chlorite begins before all the carbonates are removed.

Sample 334: Transition zone between calcareous brown (Zone IV) till and noncalcareous brown (Zone III) till. The clay minerals in this sample are intermediate in character between those of samples 335 and 333.

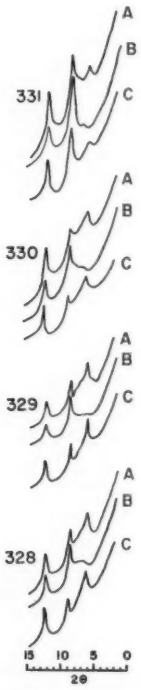

FIGURE 2.—SMOOTHED SPECTROMETER TRACES OF SAMPLES 331–328

Made on an XRD-3 General Electric recording spectrometer using filtered copper radiation. (A) sample air dried; (B) sample heated to 450°C; (C) sample treated with ethylene glycol.

Sample 333: Leached brown (Zone III) till (Fig. 1).

(1) 14 Å peak. The 14 Å peak now appears as a plateau. The shape and position of the plateau is not noticeably changed after NH_4^+ treatment and is slightly shifted to higher spacing after glycolation. The plateau is completely lost after heating to 450°C. This shows that many of the hydroxyls of the brucite sheet have been joined by protons, and the organization of the components of the brucite sheet is poor.

(2) 10 Å peak. The 10 Å peak is less intense and more asymmetrical in the untreated sample. The loss of the asymmetry after heating and the shift accompanying glycolation indicate more illite degradation and more abundant illite-montmorillonite mixed layers.

Sample 332: Extremely weathered till, Zone II (Fig. 1).

(1) 14 Å peak. The 14 Å peak in the untreated slide is a plateau as described in the previous sample. Its shape and position is relatively unaffected by NH^+ treatment, but the plateau is resolved into a 14.7 Å peak and a 12.9 Å shoulder after glycolation. These data may indicate a slight reorganization of the components of the highly hydrated brucite to form a vermiculite-chlorite mixed layer mineral.

(2) 10 Å peak. The intensity of the 10 Å peak is slightly decreased and its asymmetry increased. The loss of the asymmetry after heating and the development of the shoulder at 12.9 Å after glycolation indicate further degrading of the illite and development of more illite-montmorillonite mixed layers.

Sample 331: lower B-1 soil zone (Fig. 2).

(1) 14 Å peak. A rather broad, moderately intense peak appears at 14.3 Å in the untreated slide which is unaffected by NH^+ treatment, not completely lost but shifted to 12.8 Å after heating, and only slightly modified after glycolation. The 7 Å/14 Å peak is approximately .8/1.0. This is interpreted to mean that almost all OH^- of the brucite has been converted to H_2O, and that the H_2O and Mg^{++} are partially reorganized to form poorly crystalline vermiculite.

(2) 10 Å peak. The asymmetry of the 10 Å peak reaches a maximum in this sample. This means that the degrading of illite and development of an illite-montmorillonite mixed-layered assemblage reaches a climax in the B-1 soil zone. The ratio of illite to mixed-layer illite and montmorillonite is approximately 7/3 in this sample.

Sample 329: Upper B soil zone (Fig. 2).

(1) 14 Å peak. A sharp intense peak occurs at 14.3 Å which is not affected by glycolation or NH_4^+ treatment but is almost completely lost after heating; the 7 Å/14 Å intensity ration is .5/1.0. This indicates more vermiculite in this sample than in the preceding.

(2) 10 Å peak: A rather sharp symmetrical 10 Å peak occurs in this sample. The symmetry of the peak is little altered after glycolation and is more intense after heat treatment. An increase of the shoulder between 10 Å and 14 Å in the untreated slide and the loss of this shoulder in the glycolated and heated slide indicate illite-montmorillonite mixed layers. There are several reasons for the increase of sharpness and intensity of the 10 Å peak. The degraded illite may be regarded by K^+ re-entering the intermica layers, which thereby increases the illite crystallinity; the apparent increase of illite crystallinity may also be explained by a separation of the illite and illite-montmorillonite components into two separate phases. The

authors favor the first interpretation and suggest that potash-bearing fertilizers may be the source for the K^+ which re-enters the intermica layers.

Sample 328: A-2 soil zone (Fig. 2). The clay minerals here are similar to those of sample 329 except for a slight indication that a vermiculite-montmorillonite mixed layering may be starting to develop from the vermiculite component in this sample.

Discussion of Clay Mineral Alteration

Chlorite and Its Weathering

Chlorite alteration begins at the contact between Zone V and Zone IV. The hydroxyls of the brucite sheet are joined by protons slowly, steadily, and randomly but no "islands" of Mg^{++} regularly coordinated with H_2O develop within the brucite sheet. In Zone II of the profile almost all the hydroxyls of the brucite sheet are changed into water, and the Mg^{++} and H_2O then begin reorganizing to form vermiculite. This reorganization reaches a climax in the B-1 soil zone. The conversion of hydroxyls to water by proton addition is accelerated in the more acidified upper part of the soil profile.

The Wisconsin (Droste, 1956, p. 916) and Illinoian weathered profiles in the northwestern part of the Allegheny Plateau are alike in that vermiculite is the end product of chlorite weathering, and chlorite weathering begins in the transition zone between gray and brown calcareous till.

Although the end product of the alteration of chlorite in the Illinoian and Wisconsin profiles is almost identical, the intermediate phases are different. In the Wisconsin section the chlorite-vermiculite mixed layering begins in the lower part of the profile (Zone IV, calcareous brown till) whereas in the Illinoian section vermiculite does not appear until Zone II (thoroughly weathered till). In both profiles the hydroxyls of the brucite layer are joined by protons, but in the Wisconsin sections "islands" of Mg^{++} coordinated with surrounding water develop in the brucite sheet and produce a well-developed chlorite-vermiculite mixed layering in the lower part of the profile. No "islands" develop in the brucite layers in the Illinoian section, and vermiculite does not form until the brucite sheet is almost completely hydrated in Zone II of the weathering profile. The reason for the variation in chlorite weathering in the Illinoian and Wisconsin sections is not known; perhaps it may be explained by differences in vegetation or drainage characteristics.

Illite and Its Weathering

Degradation and hydration of the illite begins in the transition zone between gray and brown calcareous till. Probably K^+ is removed from the structure, and water enters and causes a random development of illite-montmorillonite mixed layers. This degradation and hydration continues throughout the profile with a slight reversal of the process in the highest soil zones.

The illite alteration in the Illinoian and Wisconsin profiles differs where the degrading begins. In the Wisconsin profile illite is not noticeably altered in the transition zone between gray and brown calcareous till, although in this profile illite degradation begins in this transition zone. Another difference between the profiles of the two ages occurs in the uppermost soil zones. In the Illinoian profile the illite degrading is reversed in the upper B-1 and A soil zone, and K^+ probably re-enters the intermica layers; this did not occur in the Wisconsin weathering profile.

Table 1.—Grain Size Distribution of the Weathering Profile

By percentage

Sample	Sand	Silt	Clay
328	16	48	36
329	19	56	25
330	20	53	27
331	29	46	25
332	34	38	28
333	38	38	24
334	33	37	30
335	37	35	28
336	34	36	30
337	31	36	33

Grain Size Distribution

The sand-silt-clay ratios were determined for each sample, and these data are presented to the nearest whole number in Table 1.

The average sand-silt-clay ratio of the calcareous samples is 33-36-31, and the major portion of the sand fraction is in the fine-sand size grades. This ratio is approximately the same for the samples of Zone III (sample 333) and

Zone II (sample 332) till. A marked break in the grain size distribution occurs in the lower part of the B-1 soil zone (sample 331). This break in grain size distribution corresponds with the first appearance of vermiculite as previously described.

Silty fine-grained sandstones, siltstones, and silty shales are the dominant rock types in northwestern Pennsylvania, and much of the clastic material in the till of this area is made up of fragments of these rocks. Microscopic examinations of the fine-sand and coarse-silt fractions show many sand- and silt-sized particles of siltstones and silty shales. The increase of silt and the decrease of sand beginning in the lower B-1 soil may be due to more intense weathering in the upper part of the profile which has broken down the rock fragments into smaller and even into individual grains by removing the cementing material in the rock. The increased surface area created aids the weathering of these small fragments and speeds up the alteration of the clay and nonclay minerals.

Table 2.—Clay Mineral Composition of the Weathering Profile

Till zone	14 Å mineral	10 Å mineral
Zone I	Vermiculite, no chlorite remains	More mixed layers of illite-montmorillonite, slight regrading of illite in upper part
Zone II	Strongly degraded chlorite, with a very small amount of vermiculite	More mixed layers of illite-montmorillonite; some illite unaltered
Zone III	Moderately degraded chlorite	Illite more degraded with more illite-montmorillonite mixed layers
Zone IV	Chlorite degrading begins	Illite degrading begins with a little illite-montmorillonite mixed layer development
Zone V	Chlorite	Illite

Summary

From this study it can be stated that the brucite sheet of chlorite is gradually destroyed by H^+ joining OH^- to form H_2O surrounding Mg^{++}. After destruction of the brucite layer, the hydrated components reorganize to form the vermiculite found in the subsoil. Illite is gradually hydrated, and a mixed-layer illite-montmorillonite mineral is formed. Each zone in the till profile of weathering appears to have a characteristic clay mineral assemblage (Table 2).

The sharp decrease of sand and increase of silt-size particles in the upper soil zones may be the result of more intense weathering and removal of cementing material holding the individual grains of rock fragments together.

For many years geologists and pedologists have recognized from field observations that soil profiles developed on drift of Illinoian age are different from soil profiles developed on drift of Wisconsin age. The most recent soil publication dealing with northwestern Pennsylvania (Bacon *et al.*, 1954) indicate that the soil profiles developed on the drift of the two ages are distinctly different. These differences arise from different thickness of the soil zones, different degree of structural and textural development within the soil zones, and different degree of mottling in the soil zones. Even though these differences are striking in the field, it is important to point out that the degree of weathering as shown by clay mineral alteration is essentially the same for profiles of weathering developed on tills of Illinoian and Wisconsin age.

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RECENT UNDERWATER SURVEYS USING LOW-FREQUENCY SOUND TO LOCATE SHALLOW BEDROCK

BY W. O. SMITH

ABSTRACT

Underwater investigations at Lake Mead, Chicago, Passamaquoddy Bay, and on Long Island established the characteristics of sound waves that can be used in shallow geophysical exploration by the sonar method.

At Lake Mead the sediments were for the most part clay of high water content which was easily penetrated by low-power sound at a frequency of 14.2 kilocycles. The greatest depth of penetration was 140 feet. Sound having frequencies of 50 and 80 kilocycles did not penetrate. At Chicago, sound at a frequecy of 11 kilocycles and an output power of 800 watts gave a satisfactory delineation of bedrock beneath Lake Michigan. The maximum distance to bedrock was about 135 feet. At Passamaquoddy Bay a sound frequency of 6 kilocycles and about 700 watts of output power gave much better delineation of bedrock. A frequency of 6 kilocycles at the lower output power was much better than 11 kilocycles at higher power. About 250 feet of penetration was attained. Pulsed power was used in each of these investigations. The pulse lengths were long—about 14–25 milliseconds.

Bedrock was mapped at Lake Mead, Chicago, and Passamaquoddy Bay. The methods of ordinary hydrographic surveying were used for horizontal and vertical positioning. For horizontal positions the ordinary three-point sextant-fix method was used. For vertical positioning, recording gages suitably placed and supplemented by staff gages were used. All data were tied into the third-order control net of the U. S. Coast and Geodetic Survey.

In connection with a study for a proposed Midwestern waterway, the technical problem was reviewed and the techniques considerably improved. New equipment was built and evaluated on Long Island Sound. It operated with pulsed power at a frequency of 6 kilocycles, and pulse length was controlled and variable from 1 to 9 milliseconds. Output acoustic power was about 2500 watts. With the transducer in ordinary operating positions as much as 400 feet of sediment was penetrated. With the transducer placed directly on the bottom of the water in Huntington Bay about 750 feet of penetration was attained. Several innovations in sonar techniques, which are desirable for sediment exploration, and which give much more detailed information than the earlier equipment are described.

The techniques required for stratigraphic interpretation of the sound records are described briefly. A method for the determination of sound velocities is discussed. Multiple echoes and other effects complicate the interpretation of records and are explained. The problems of geological control encountered also are mentioned.

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Introduction

During the Lake Mead Comprehensive Survey of 1947–1949, it was observed (Smith *et al.*, 1954) that low-frequency sound penetrated substantial thicknesses of sediment in Lake Mead. Subsequently, a further opportunity to investigate the observed phenomena was presented by two projects: first, at Chicago where bedrock elevations were determined for use in the design and placement of the Central District Filtration Plant and its proposed tunnel system; and second, in a study of bedrock beneath areas in Passamaquoddy Bay being considered as dam sites for a proposed tidal-power project. Assisting with plans for a Midwestern waterway investigation afforded an opportunity to study the problem still further, in the light of what had been learned at Lake Mead, Chicago, and Passamaquoddy Bay. Equipment specifically suited to the purpose was designed and built. Detailed tests were made on Long Island Sound.

This paper gives a brief account of each of the investigations and emphasis is on those characteristics of sound that relate to penetration in sediments—specifically, sound-wave frequency, pulse length, and output power level. Some consideration is given also to receiver sensitivity and its effect upon the recording of useful data on sediment and rock depths.

A relatively detailed account of the Long Island Sound investigation is given, because on that expedition many fundamental data were collected to illustrate graphically what can be done with high-power, low-frequency, pulsed sonar.

The equipment used in each of these investigations consisted essentially of the same basic elements. A transducer served to convert electrical energy of known frequency into acoustic energy and to direct it outward into the water and sediment. Pulsed power, with a variable pulse length, was used. Generally, the same transducer picks up the returning sound echo and reconverts it into electrical energy. Pulsed power is desirable in sonar work because it facilitates a direct measurement of time—specifically, the interval lasting from the

moment a given sound wave leaves the equipment until it returns after reflection. This is a fundamental principle of sonar when pulsed power is used. Depths of penetration are calculated from sound velocities determined in the media involved—water, sediments, or consolidated rock.

The electrical energy for the outgoing pulse is supplied as follows: A signal of desired frequency is generated in a suitable oscillator and, after it is keyed and spaced to get the desired pulse length and duty cycle, it is passed to a power anplifier and finally through a matching network to the transducer. The returning echo, after being converted in the transducer to electrical energy, passes through a receiver where it is detected and amplified; the output passes to a recorder where a permanent record is made, or else to a cathode-ray tube for visual examination. Much detailed circuitry and intricate mechanical design are required to produce satisfactory equipment. These will not be discussed at length in this paper, since the emphasis is to be on the geophysical aspects of the problem.

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The writer wishes to thank George B. Cummings of the Telephonics Corp., and Charles E. Mongan, formerly of the Edo Corp., for many helpful suggestions. The assistance of Joseph E. Upson, staff geologist of the Geological Survey, Mineola, N. Y., is greatly appreciated.

Lake Mead Investigations

During the Lake Mead Comprehensive Survey of 1947–1948 (Smith *et al.*, 1954, v. 2, p. 110–113), it was observed that low-frequency sound penetrated substantial thicknesses of unconsolidated lake-bottom sediments. An examination of the sound records showed complete penetration of existing sediments over much of Lake Mead between Hoover Dam and the Virgin River. Most of the sediment deposited to date is confined to the areas in or near the former Colorado and Virgin River channels.

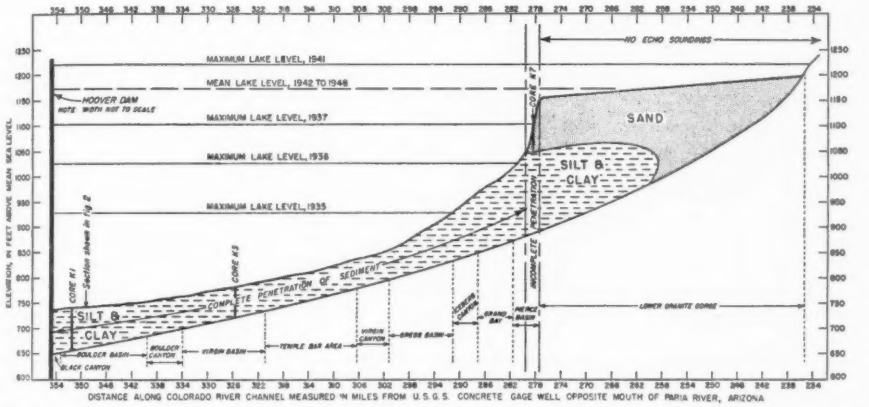


FIGURE 1.—PROFILE OF SEDIMENT DISTRIBUTION IN COLORADO RIVER CHANNEL FROM HOOVER DAM TO THE END OF THE LOWER GRANITE GORGE, SHOWING SOUND-PENETRATION DATA

The depth of sediment in Lake Mead in 1948–1949 ranged from nothing in certain parts of the Overton Arm to approximately 275 feet in the lower Granite Gorge. The sediment near Hoover Dam was about 140 feet thick, and it thinned to about 80 feet near the eastern reaches of Boulder Basin. It was about 60 feet thick in the center of Virgin Basin. It then thickened again eastward along the old river channel in Gregg Basin and Grand Bay, and also in Pierce Basin, as the lower Granite Gorge was approached. Figure 1 is a thalweg profile of sediment distribution from Hoover Dam to the end of the lower Granite Gorge. It follows the old Colorado River channel passing through the principal lake areas, Boulder Basin, Virgin Basin, Gregg Basin, Grand Bay, Pierce Basin, and the lower Granite Gorge.

Complete acoustic penetration of the sediments was obtained throughout the deep leak from Boulder Basin to Grand Bay and perhaps just into the western part of Pierce Basin.

TABLE 1.—TYPICAL CORE DATA FOR LAKE MEAD

Depth Below Sediment Surface (Feet)	Water Content by Weight (Per Cent)	Wet Density (g/cc)	Specific Weight (lbs./cu. ft.)	Porosity	Water Depth (Feet)	Sediment Thickness (Feet)
Core K-1—Boulder Basin: Location Lat. (N.) 36° 02.47′ Long. (W.) 114° 44.43′ River Mile 351.80						
7.8	65.4	1.275	27.6	83.5	409.2	83.7
12.4	64.0	1.284	28.9	82.9		
20.5	63.3	1.295	29.7	82.4		
20.9	60.9	1.314	32.1	81.0		
32.2	57.3	1.367	36.4	78.4		
39.0	52.2	1.429	42.7	74.7		
47.2	46.8	1.503	49.9	70.4		
50.6	43.4	1.553	54.9	67.5		
53.1	40.9	1.591	58.6	65.2		
54.7	33.5	1.718	71.4	57.7		
Core K-3—Virgin Basin: Location Lat. (N.) 36° 07.58′ Long. (W.) 114° 26.65′ River Mile 325.0						
3.9	59.5	1.329	33.6	80.0	368.0	48.8
7.0	61.9	1.311	31.2	81.6		
11.1	59.6	1.340	33.8	80.0		
18.0	57.3	1.368	37.4	78.4		
20.9	52.5	1.426	42.3	74.9		
26.7	51.9	1.433	43.0	74.5		
31.5	43.3	1.551	54.9	67.5		
32.9	43.8	1.540	54.0	68.0		
39.0	34.5	1.699	69.5	58.9		
44.5	41.9	1.570	56.9	66.3		
Core K-7—Pierce Basin: Location Lat. (N.) 36° 07.92′ Long. (W.) 113° 58.68′ River Mile 279.29						
0.9	37.8	1.638	63.7	62.3	91.5	156.8
8.0	28.9	1.809	80.3	52.5		
12.7	20.9	1.959	96.8	42.7		
17.0	36.4	1.650	65.6	61.3		
26.2	21.7	1.971	96.4	43.0		
33.3	23.6	1.923	91.8	45.7		
36.0	23.6	1.909	91.0	46.1		

Within Pierce Basin, however, little, if any, penetration was found. No echo soundings were taken in the lower Granite Gorge. The regions of complete and incomplete penetration are shown in Figure 1.

The question arises as to why there is a difference in acoustic penetration. The answer is not simple and seems to depend upon both the nature of the sediments and the sound characteristics. It will be seen from Figure 1

DEPTH, IN FEET BELOW LAKE STAGE AS OF APRIL 27, 1948

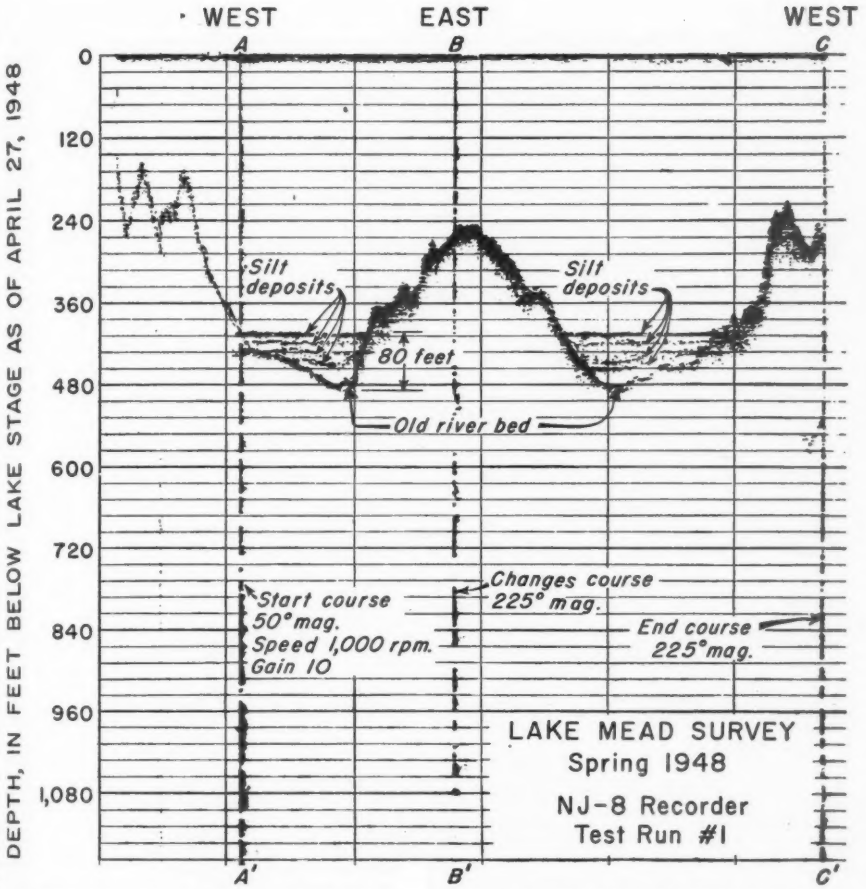


FIGURE 2.—SOUND RECORD IN BOULDER BASIN, ACROSS FORMER COLORADO RIVER CHANNEL
(Location is shown in Fig. 1)

The part of the record included between fix lines *A-A'* and *B-B'* was run from the Nevada to the Arizona shore, and that part included between fix lines *B-B'* and *C-C'* is approximately the same course run in the reverse direction. The three dark lines between the surface of the silt deposits and the old river bed presumably are contacts between seasonal density flows.

that in the regions of complete penetration the sediments consist for the most part of silt and clay. Near Hoover Dam they are chiefly clay, but in Pierce Basin and eastward they are chiefly sand. Also, the sediment thickness is very much greater in Pierce Basin.

Table 1 shows the water content, specific weight, and porosity of core samples at river mile 351.8 in Boulder Basin, mile 325 in Virgin Basin, and mile 279.3 in Pierce Basin. The location of each is shown in Figure 1.

The cores were taken with a piston-type core sampler of the Kullenberg type (Kullenberg, 1947), which had been considerably modified by the author for the work in Lake Mead. The cores appeared to be quite good, with little or no observable shortening. The core barrel of the sampler consisted of 33 sections of hardened copper pipe joined together by tapered screw connections. Each section was 3 feet long and 2 inches in diameter. The core segment in each 3-foot section was broken into shorter parts for laboratory examination by pressing it, with a piston, through a special adapter in ⅛-inch-wall

lucite tubes, each 2 inches in diameter and 6 inches long. The tubes were then sealed.

The determinations given in Table 1 (Smith *et al.*, 1954, Supplemental Base Data Report) were made immediately after the taking of each core. The specific weights of the samples taken in Boulder and Virgin Basins are noticeably lower than those taken in Pierce Basin, where almost no penetration occurred. Likewise, the water contents are considerably higher in the samples from Boulder and Virgin Basins than they are in Pierce Basin. The conclusion seems to be that, as far as the physical nature of the sediments in Lake Mead is concerned, complete penetration was secured in sediments consisting of clays of high water content, whereas little or no penetration occurred in sands of low or moderate water content. The water content and texture of the sediments are, however, not the only factors controlling the degree of sediment penetration. The characteristics of the sound wave also are important.

Three sound frequencies, 80, 50, and 14.2 kilocycles, were used in the Lake Mead investigation (Smith *et al.*, 1954, v. 2, p. 110–113). At no time was even the slightest penetration observed at either 80 or 50 kilocycles, regardless of the type of sediment. It occurred only with the frequency of 14.2 kilocycles. The fact that only low frequencies will penetrate sediments is not surprising. Acoustic waves, like compressional seismic waves, are compression waves in matter. The 14.2-kilocycle sound frequency is near the upper limit o f the usable seismic band

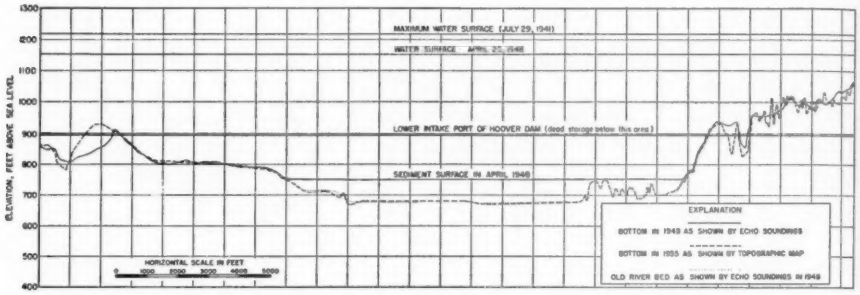


FIGURE 3.—PROFILE OF OLD COLORADO RIVER CHANNEL IN BOULDER BASIN

(Location is near section shown in Fig. 2) The profile obtained from sound data in the 1948–1949 survey is compared with that taken from the topographic map made before the lake filled.

It is to be observed that, as in seismic work, the lower frequencies are generally more satisfactory and penetrate with less attenuation. The output acoustic power at Lake Mead was low—of the order of 20 watts at each of the frequencies tried. The 80-kilocycle and 50-kilocycle frequencies gave a much sharper delineation of the top of the sediment than did 14.2 kilocycles. The 14.2-kilocycle frequency, however, was satisfactory in delineating the top of the sediment.

Figure 2 is a photograph of a sound record across the old river channel in Boulder Basin. A maximum penetration of 80 feet is shown. Several bands of sediment are observable; each one possibly represents a seasonal density flow. The sediments shown have a relatively high water content.

Figure 3 shows comparative data on the sediment bottom. Data obtained solely from sound records are compared with those obtained from the topographic map (Brown, 1940—1941)

PLATE 1.—SOUND RECORDS IN LONG ISLAND SOUND JUST NORTH OF MANHASSET BAY

FIGURE 1.—Comparative sound records over test site near fix 10, on range *AA*, north of Manhasset Bay. 2-millisecond pulse lengths are used. (a) Sound record at 11 kilocycles (b) Sound record at 6 kilocycles.

FIGURE 2.—Sound records near fix 3, range *AA* (Fig. 18) just north of Manhasset Bay (a) The overburden is well developed; the bedrock shows but is weak (b) The bedrock is clear, but just over the bedrock the overburden is not well developed.

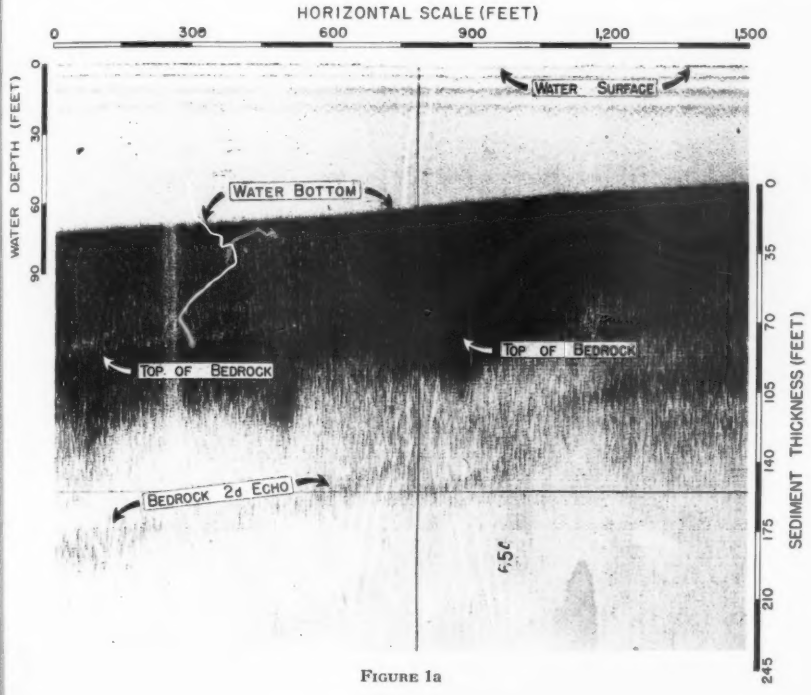


FIGURE 1a

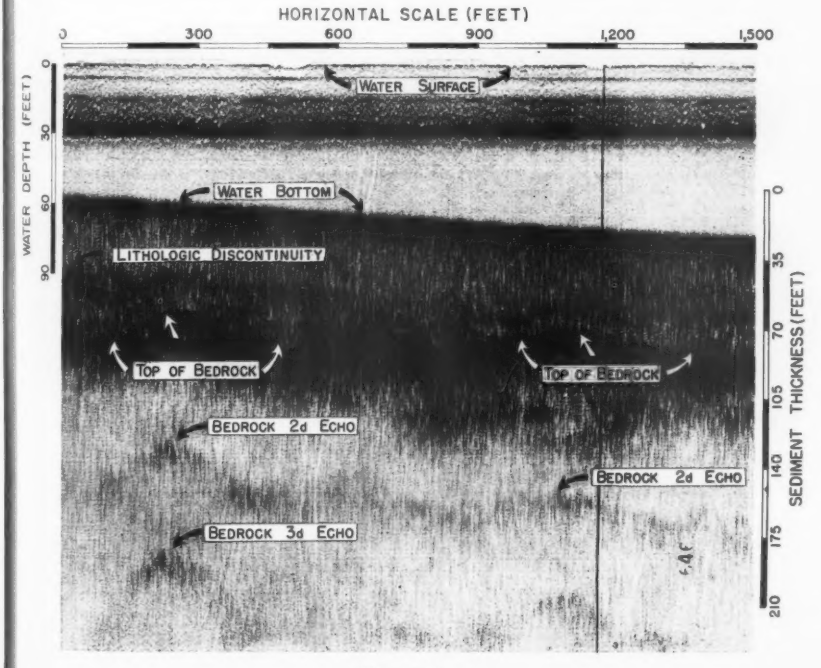
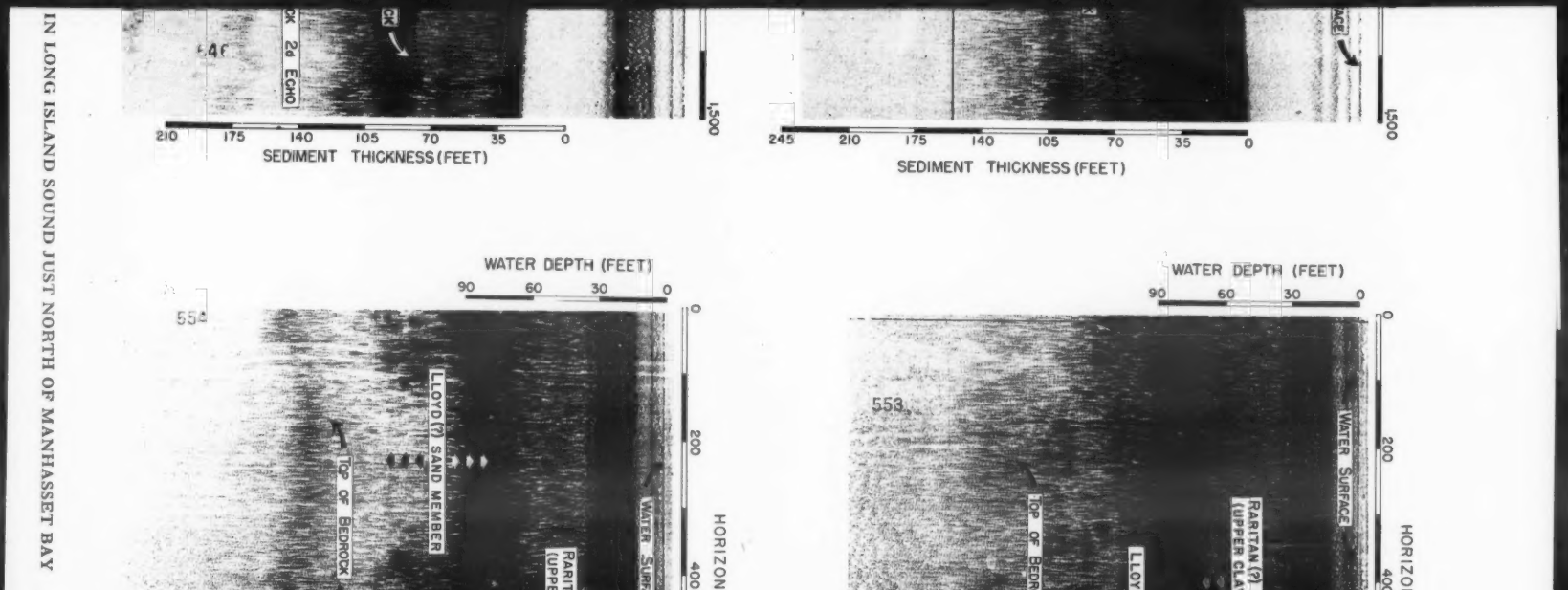


FIGURE 1b



IN LONG ISLAND SOUND JUST NORTH OF MANHASSET BAY

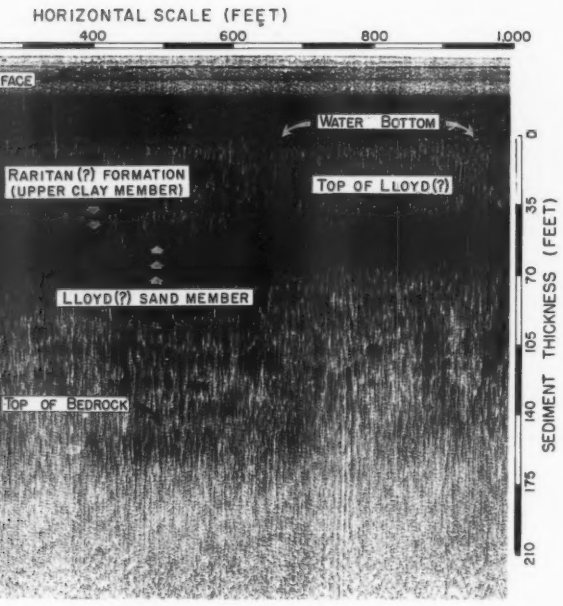


FIGURE 2a

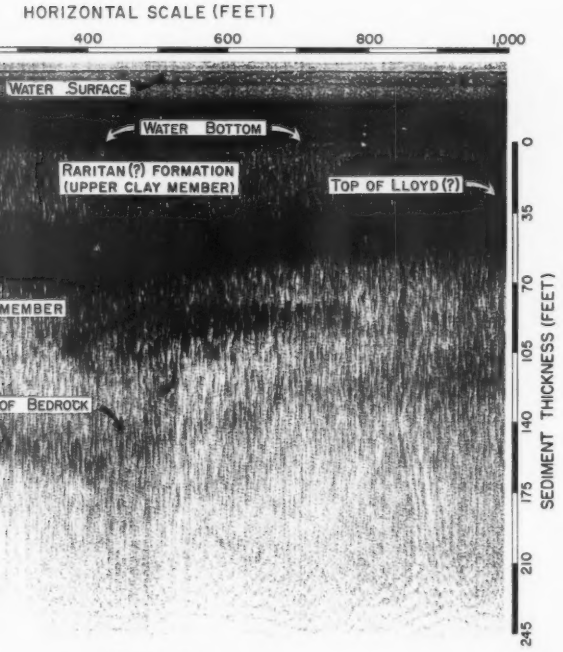
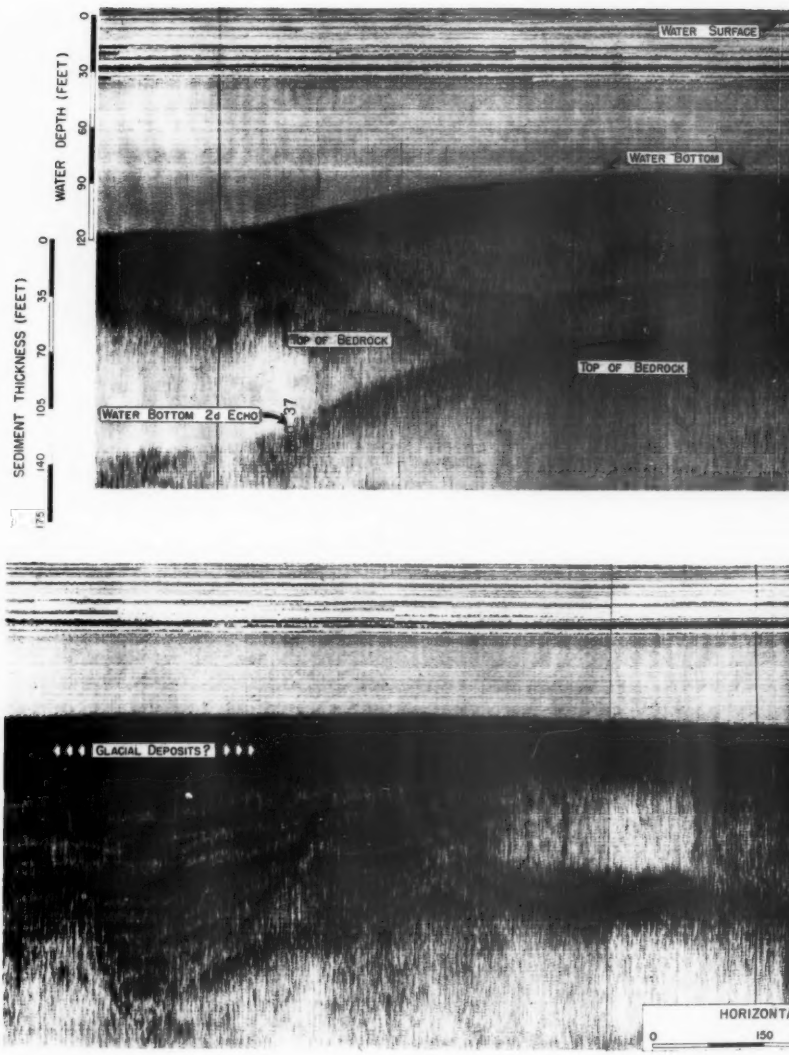


FIGURE 2b

SET BAY



SOUND RECORD ALONG NORT
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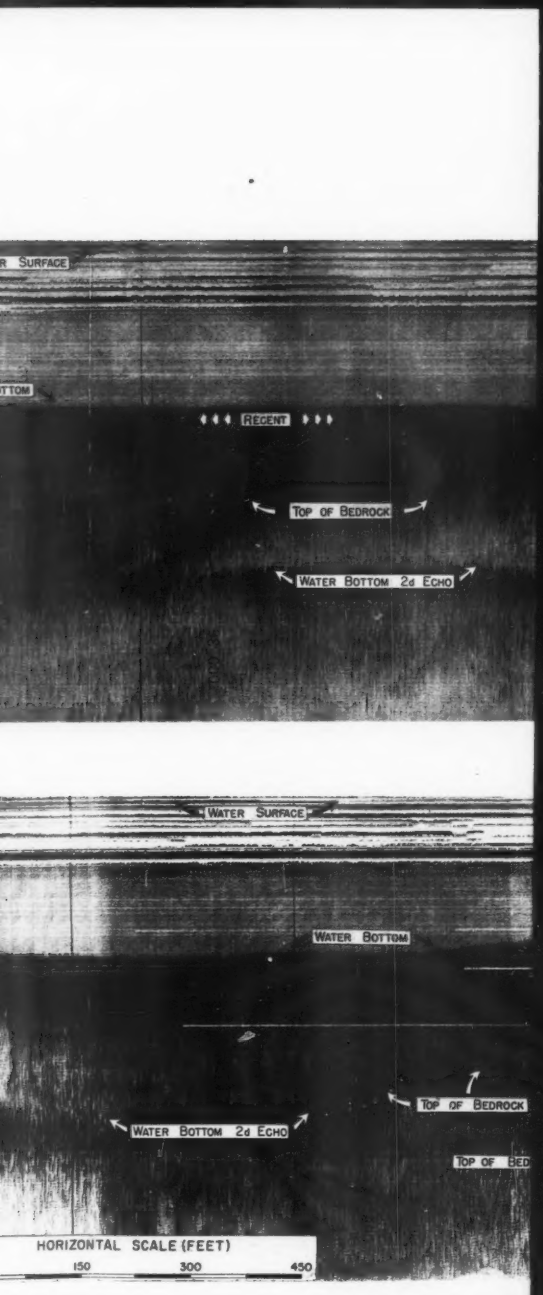
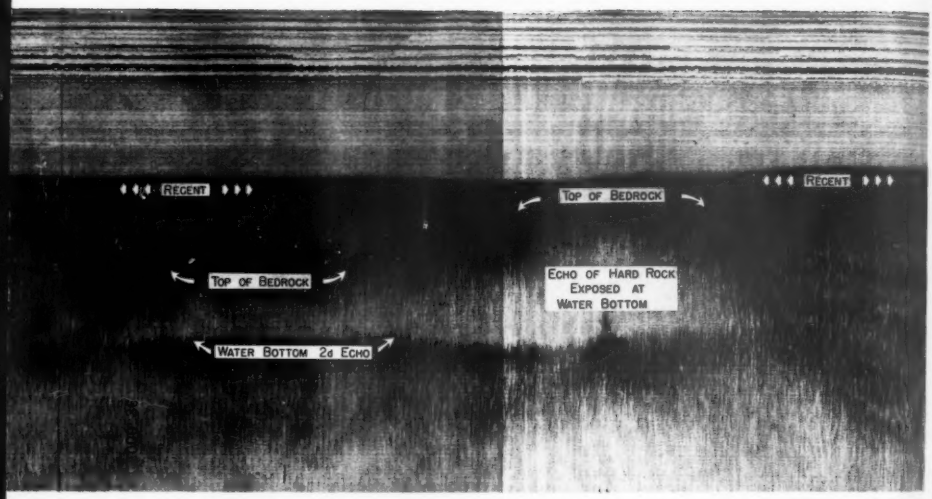
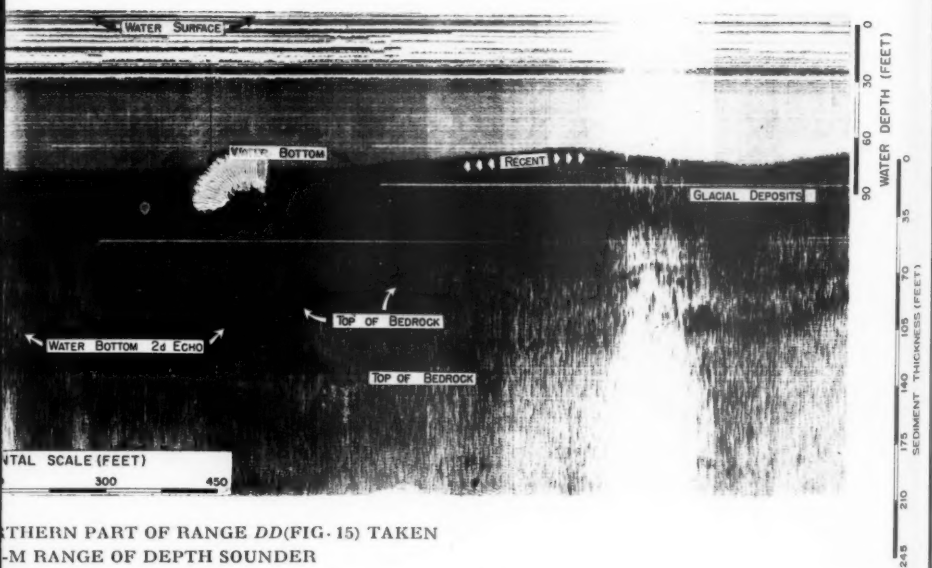


ONG NORTHERN PART OF RANGE *DD*(FIG. 15) TAKEN
E 0- TO 70-M RANGE OF DEPTH SOUNDER





THERN PART OF RANGE *DD*(FIG. 15) TAKEN
-M RANGE OF DEPTH SOUNDER

made in 1935 before the lake was filled. The sound velocity in the sediment is taken as 5000 feet per second. Good agreement is shown between the two sets of data. The section shown in Figure 3 is in Boulder Basin and close to that shown in Figure 2.

The velocities of sound were ascertained in sediments by methods described under "Determination of sediment depths and sound velocities." The determination was particularly simple in the Lake Mead case. A few well-chosen locations were sufficient.

The sound velocity in water was determined first from measurements of salinity and temperature, and second from actual calibration with a check bar. Both methods are common in hydrographic surveying. In the first, tables giving the value of sound velocities in water as a function of salinity and temperature were used. In the second, a check bar was placed at a known depth beneath the water surface, and the time of travel from the transducer to it was determined directly from the echo-sounder chart and the constants of the recorder. Since the depth was known, the velocity was easily calculated. If the sound chart recorded depths, corrections were easily determined. Thus the elevations of the water bottom and top of the sediment became known.

To determine the sound velocity in sediments at a chosen location, both the sediment thickness and the time of travel of the sound wave from the top to the bottom of the sediment had to be known. The sediment thickness was known if the elevations of its top and bottom were established. In the Lake Mead work the elevation of the top of the sediment was determined from the sound charts by the procedure previously described. The elevation of the sediment bottom was determined from topographic maps made in the areal survey in 1935. The nature of the rocks on which the sediment rests was known from the geologic study made by Longwell (1936) before filling of the reservoir. The sediment bottom was easily recognized on the sound charts, as illustrated in Figure 2. The time of travel of the sound wave from the top to bottom of the sediment could be determined from the sound-chart reading and the constants of the recorder. Thus the velocity of sound in the type of sediment at a specified location could be easily calculated. Only a few locations were used in each area. Sound velocities used to calculate the profile of the old river bed shown in Figure 3 were an average of several others, east and west of the section.

The velocity of sound where penetration occurred in sediments of Lake Mead was generally about 5000 feet per second, about the same as that in water. This was not surprising, since the deposits in Lake Mead are silt and clay of high water content, and hence rather fluid. Some gas also is entrapped.

The Lake Mead studies established with certainty that low-frequency sound would penetrate a moderate thickness of fine-grained sediments of high water content. Although the power output of the sonar equipment was low, the sound level was adequate for penetration and gave a recordable return echo, presumably because of the fluidity of the sediments.

Determination of Sediment Depths and Sound Velocities

In order to understand the Chicago and subsequent investigations, some knowledge of how sound velocities are determined in sediments is essential. Hence a simple account of a basic approach to the problem will be given before proceeding to a discussion of the further investigations.

To determine the depth to a given reflecting surface, the velocity of sound in the intervening medium in which the sound wave travels must be known. These determinations are relatively simple when pulsed power is used. The processes involved are illustrated in Figures 4*a* and 4*b*. Figure 4*a* is an illustration of the process in water. The outgoing pulse P_w leaves the transducer T, located at a distance S_t below the water surface, and travels to the water bottom Z_w along the path A. A part of its energy is reflected at the water bottom and forms the return echo E_w, which returns to the transducer along the path B. If the time of travel from the transducer to the water bottom is T_w, the total time of sound travel from the transducer to the water bottom and return is 2 T_w. The sound chart records this value. If S_w is the distance from the water surface to the water bottom and S_t is the distance from the water surface to the transducer, then, if V_w is the velocity of sound in water,

$$S_w - S_t = V_w T_w \qquad (1)$$

V_w = 4800 ft./sec. for sea water at a temperature of 41°F. and a salinity of 3.2 per cent. For other salt concentration and temperatures the corrections are to be found in tables. Cor-

rections for the depth of the transducer below the water surface are required for accurate depth determinations. Using the fixed value of V_w for water, suitably corrected for salinity and temperature, recorders may be calibrated directly in water depths.

If T_w is the time of travel for the distance $S_w - S_t$ from the transducer T to the water bottom Z_w, the sound recorder records the time $2\,T_w$ and equation (1) holds as before.

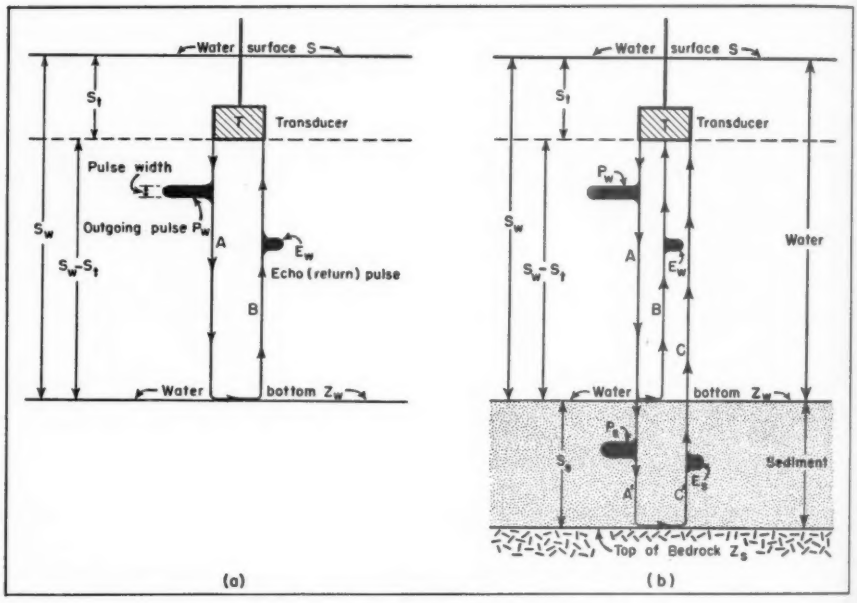


FIGURE 4.—TRANSMISSION OF SOUND PULSES

(a) In water as related to the transducer, water surface, and water bottom; (b) and in a simple sediment as related to the transducer, water surface, water bottom, and sediment bottom.

Figure 4*b* illustrates the reflections in a single sediment layer overlain by water. The sound pulse P_w leaves the transducer T and travels to the water bottom Z_w along the path A. At Z_w it divides. A part is reflected as the echo E_w from the water bottom Z_w and returns to the transducer along the path B. The other part of the pulse P_w continues on through the sediment as the pulse P_s along the path A' to the sediment bottom Z_s. Here it is reflected as the echo E_s returning along the paths C' and C through sediment and water respectively back to the transducer. It should be observed that the travel times for equal distances in water and sediment are not the same because the sound velocities are generally faster in sediments than in water. One exception is gas-bearing sediments, in which the sound velocity may be less than in water.

If T_s is the time required for sound to travel from the water bottom Z_w to the sediment bottom Z_s, then $2\,T_s$ is the value indicated on the recorder. From Figure 4*b*

$$T_s = S_s/V_s \tag{2}$$

The time of travel T_R of the sound wave from the transducer to the reflecting sediment bottom is given by $T_R = T_w + T_s$, and

$$T_R = (S_w - S_t)/V_w + S_s/V_s \tag{3}$$

which is sufficient for calculating the sediment velocity V_s if the other quantities are known. For the purpose of calibration the distance S_s must be obtained from known data, preferably a sound record which has been run over a previously logged borehole to find the corresponding value of T_s.

The argument can be extended to sedimentary strata consisting of several different types

Suppose that, in addition to water, there are several sedimentary strata lying between the transducer and a given reflecting rock surface. If V_W, V_{R_1}, V_{R_2}, V_{R_3}, ... V_{R_N} are the velocities of sound in water and in the rock strata R_1, R_2, R_3, ... R_N, and if T_W, T_{R_1}, T_{R_2}, T_{R_3}, ... T_{R_N} are the times of travel to the water bottom, and within the successive single strata, and if, further, S_W is the depth of water, and S_{R_1}, S_{R_2}, S_{R_3}, ... S_{R_N} are the respective thicknesses of successive rock strata, then if S_{R_N} is the distance from the transducer to the bottom of the reflecting rock strata R_N, we have

$$S_{R_N} = V_W T_W + \sum_{i=1}^{N} V_{R_i} T_{R_i} \qquad (4)$$

If, however, as often happens in simple sediments where the velocity of sound does not vary by more than a few per cent from stratum to stratum, an average value $\bar{V}_R$ may be used, and

$$S_{R_N} = V_W T_W + \bar{V}_R T_{R_N} \qquad (5)$$

may be written where T_{R_N} is the time of travel from the water bottom to the reflecting stratum. If, for example, the sound velocity varies from 6200 ft./sec. to 6500 ft./sec. in unconsolidated sediments in the first 500 feet of overburden (measured from the water bottom), and a mean value $\bar{V}_R$ of 6350 ft./sec. is used, the error is of the order of 2 per cent. This would mean 2 feet in each 100 feet of overburden. If extreme accuracy is required, then the velocity of sound for each stratum must be determined from sound-time data observed at places for which reliable well logs are available, and equation (4) used to calculate the depths of the strata. The velocities V_{R_1}, V_{R_2}, V_{R_3}, ... V_{R_N} previously defined are required. They are best found from a set of simultaneous equations derived from equation (4) and data from known drill holes.

Simple considerations, such as the above, are sufficient for calculating sound velocities from equations (1) to (5). When sound velocities have been determined, sediment thickness can be calculated from equation (4) or (5). The problem will be illustrated further in connection with the account of the Passamaquoddy investigation. Recent seismic work by Ewing (1950) and Oliver and Drake (1951) on the Atlantic continental shelf is very helpful, since the seismic velocities are essentially the same as the acoustic velocities.

Multiple Echoes

In general it is not difficult to select true rock echoes. Multiple echoes must be considered. They arise in sediments as well as in water. The effect in sediments is well illustrated by considering a simple case where a water surface, a water bottom, and an underlying bedrock exist. The overburden can be considered homogeneous. The simplest case is that of reflection of a part of the first bedrock echo at the water surface, and its return to the water bottom, where it is reflected to form a second bedrock echo. Another case is that arising from return from the water surface of a part of the energy of the first bedrock echo back through the water bottom and its second reflection at the top of the bedrock, back through the water bottom and to the transducer. This is another kind of second bedrock echo. Very unusual conditions can arise from internal reflections within complex sediments. The spacing problem is not as simple, because sound velocities are different in sediments and in water. The analysis will not be developed further in this writing.

Chicago Investigations

An opportunity for additional study arose in July 1950, particularly with reference to the effect of frequency and output power. The City of Chicago, in connection with its proposed Central District Filtration Project (Salter, 1956), requested the Geological Survey to participate in a cooperative study to map the bedrock just offshore in Lake Michigan, using sonic means.

A preliminary study of existing cores from the area was made. It indicated, among other things, that the water content was much less than in the cores from Lake Mead. The use of the low-frequency Lake Mead equipment for this study was considered to be questionable, since it had a low power output and a moderately low frequency. Tests at Lake Michigan confirmed these doubts. The high-frequency equipment, however, was considered satisfactory to map the top of the sediment and was later used for this purpose.

The Navy Department's Bureau of Ships had just developed depth-finding equipment of much greater power output, operating at a frequency of 11 kilocycles. It used pulsed power with pulses of about 20-millisecond duration. It was felt that, with this equipment, there would be a fair chance to penetrate the sedi-

ments of Lake Michigan and reach bedrock. The equipment was made available to the Survey, and the proposed investigation was undertaken.

Standard hydrographic-surveying methods were used for control. The area investigated was roughly 4 miles wide and 6 miles long, a strip extending from Navy Pier to the Wilson Avenue tunnel line. Suitable topographic control was established and tied into a standard reference net over the area to be mapped, shown in a location map of the Chicago area, Figure 5. A course of appropriate range lines was then laid out. Good geologic control existed. (For accounts of the areal geology *see* Horberg, 1942, unpub. Ph.D. thesis, Univ. Chicago; 1950; Taylor, 1930, unpub. Ph.D. thesis, Northwestern Univ.) There were many borehole logs, from which an accurate map of the bedrock under the land area was prepared. From these data, both borehole and sonic, two maps were drawn; one showed the elevation of the top of the sediment, and the other showed contours on bedrock. The offshore bedrock topography as developed from underwater sound data tied in closely with those of the shore areas developed from well logs.

Sound velocities were determined by the procedure outlined earlier using equations (1) and (3). They were established by running sound recordings at the locations of two sets of logged underwater borings, one at the north end of the area surveyed and one at the south end. Six borings were in the north along the line of the Wilson Avenue tunnel system shown in Figure 5. The other set was near Navy Pier at the site of the proposed filtration plant. The two sets of boreholes furnished the geologic control in the lake.

Some difficulty with reverbation effects was encountered in shallow water near shore when the transducer, or sound head, was in normal position near the water surface. It was important to have good sound records in these locations since they furnished the tie in with the shore-bedrock data. In order to meet this difficulty, John Campani, a sound engineer in the party, suggested that the transducer be placed on the water bottom, directly on the sediments. Thus no bulk water mass existed between the transducer and overburden. His suggestion was followed, and a much sharper echo was obtained. The effect is probably due to better acoustic coupling to the sediments when the transducer rests directly on them, as well as to the absence of blanketing reverberation, so troublesome in shallow water. To reduce reverberation still further the entire transducer, with the exception of the transmitting diaphragm, was wrapped with special heavy sponge rubber. It did not become water soaked, and it is unlikely that its air cells collapsed to any degree at the water depths where it was used.

The transducer was placed on the bottom at 22 intermediate locations (Fig. 5). These points were selected to be as near known boreholes as possible, or near contours drawn on the basis of logs of borings on shore. These data were used chiefly to supply close-in, offshore geologic control and background for interpreting bedrock traces on the sound record when the transducer was in its normal operating position near the water surface farther offshore. The records were of great value for this latter purpose, especially where the offshore sound records were feeble.

The locations of the boreholes along the Wilson Avenue tunnel system used for geologic control are shown in Figure 6. The texture of the overburden and the bedrock section also are shown in this figure. Figure 7 locates the boreholes used for geologic control in the Navy Pier area. Texture and water contents are

PLATE 3.—TYPICAL SOUND RECORDS FROM CHICAGO, AND FROM PASSAMAQUODDY BAY AREAS

FIGURE 1.—Sound record near Carter-Harrison crib (Chicago Area). The water bottom, bedrock, and some intermediate sediment contacts are shown.

FIGURE 2.—Sound record, in Belmont Harbor (Chicago) with transducer directly on the water bottom. Bedrock and intervening contacts, and second echoes of each are also shown.

FIGURE 3.—Sound record in the Pope Island-Deer Island area of Passamaquoddy Bay. A contact between clay and a sand and gravel deposit can be seen in addition to bedrock.

PLATE 4.—SOUND RECORDS ON THE 0- TO 140-M RANGE OF DEPTH SOUNDER

FIGURE 1.—Sound record along northern part of range *DD* (Fig. 15).

FIGURE 2.—Transducer on the water bottom; effect of output power level. The effect of TVG (time varied gain) is also shown.

FIGURE 1 FIGURE 2 FIGURE 3

TYPICAL SOUND RECORDS FROM CHICAGO AND FROM PASSAMAQUODDY BAY AREAS

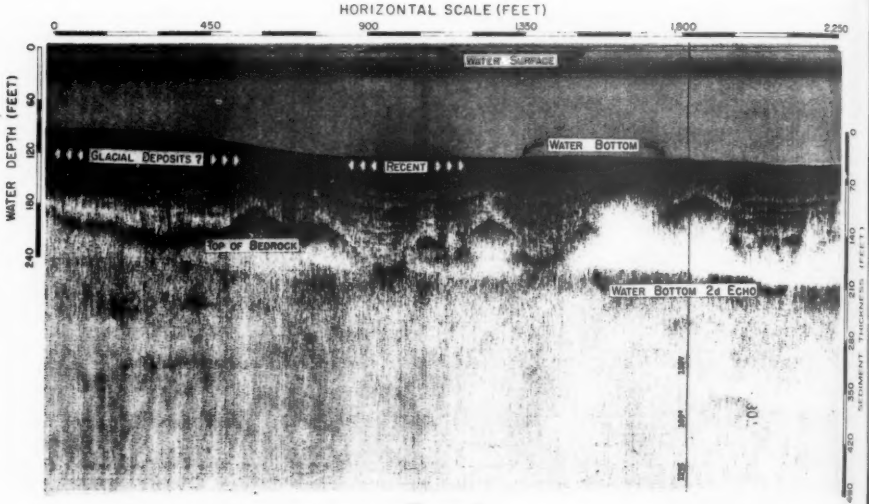


FIGURE 1

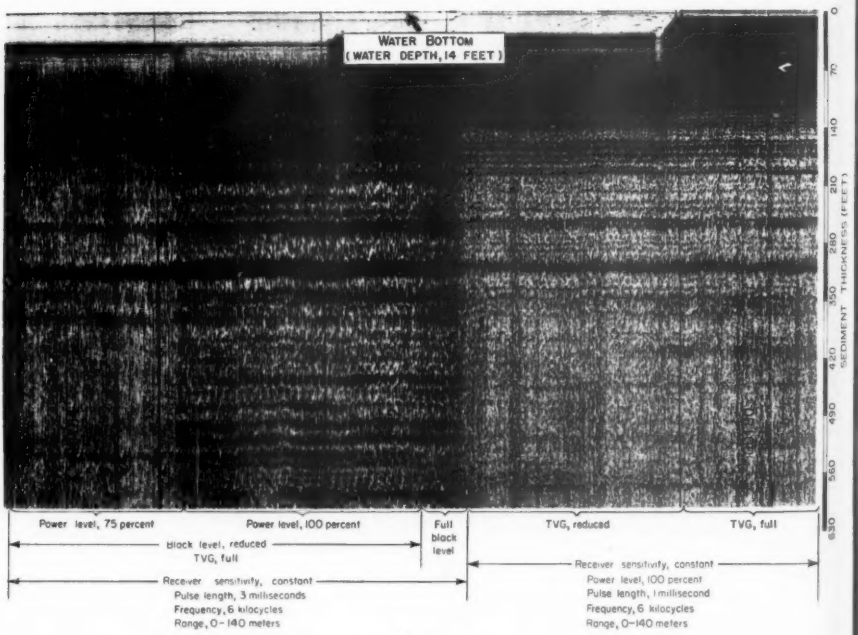


FIGURE 2

SOUND RECORDS ON THE 0- TO 140-M RANGE OF DEPTH SOUNDER

EFFECT OF COUPLING

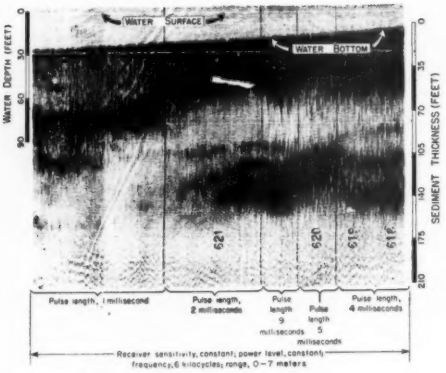


FIGURE 1

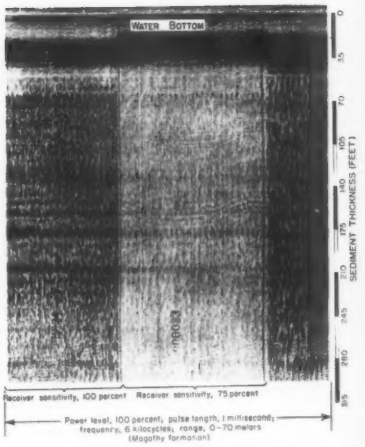

WATER BOTTOM
SEDIMENT THICKNESS (FEET)
Receiver sensitivity, 100 percent
Receiver sensitivity, 75 percent
Power level, 100 percent; pulse length, 1 millisecond; frequency, 6 kilocycles; range, 0–70 meters
(Magothy formation)

FIGURE 2

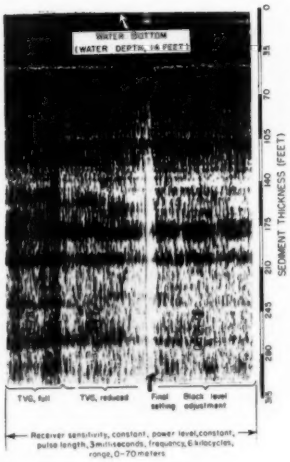

WATER BOTTOM
(WATER DEPTH, 14 FEET)
SEDIMENT THICKNESS (FEET)
TVG, full
TVG, reduced
Fixed setting
Black level adjustment
Receiver sensitivity, constant, power level, constant, pulse length, 3 milliseconds, frequency, 6 kilocycles, range, 0–70 meters

FIGURE 3

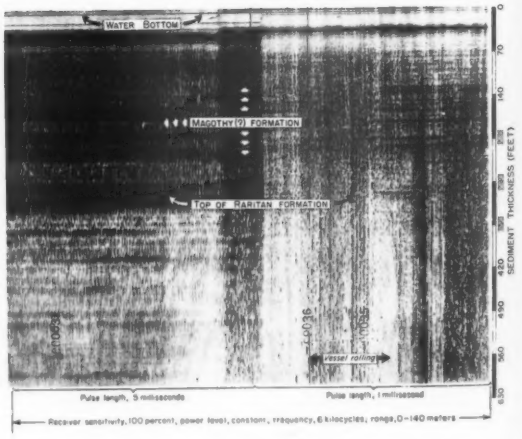

WATER BOTTOM
MAGOTHY(?) FORMATION
TOP OF RARITAN FORMATION
SEDIMENT THICKNESS (FEET)
Vessel rolling
Pulse length, 5 milliseconds
Pulse length, 1 millisecond
Receiver sensitivity, 100 percent, power level, constant, frequency, 6 kilocycles, range, 0–140 meters

FIGURE 4

EVALUATION OF SONAR PARAMETERS

shown in Figure 8 for two boreholes in the geologic-control area near Navy Pier. The sediments penetrated consist largely of clay, some fine sand and gravel, and some sand and gravel with an underlying hardpan. The water contents of the overburden are noticeably lower than those of the very recent deposits in Lake Mead.

Figure 1 of Plate 3 shows an 11-kilocycle sound record for a site near Carter-Harrison crib. The water bottom, bedrock, and some intermediate sediment contacts are shown. Figure 2 of Plate 3 shows an 11-kilocycle sound record in Belmont Harbor, with the transducer directly on the water bottom. Bedrock and intermediate sediment contacts are shown. Second echoes of each also can be seen. These are not the ordinary multiple echoes which arise from reflections between the water surface and the bottom, because the transducer is directly on the bottom. Since the pulse lengths are of the order of 14–20 milliseconds there is some overlapping. Most of the sound records obtained with the transducer in normal operating position were comparable to those shown in Plate 3, Figure 1 but some were weaker. A few were not readable. Since there were not enough of these to interfere with the bedrock mapping, and preservation of operating schedules was necessary, no attempt was made to learn the cause of the difficulty.

Figure 9 shows a part of the bedrock map finally developed. The location is in the vicinity of Lincoln Park. The offshore contours based on sonic data tie in satisfactorily with those on shore, based on well logs.

Passamaquoddy Bay Survey

Another bedrock exploration was made in Passamaquoddy Bay in August 1951, in connection with a tidal-power project under consideration for that area. This investigation was undertaken in cooperation with the New England Division of the Corps of Engineers, U. S. Army. Here the purpose was to map bedrock and its overburden in certain regions selected with regard to their suitability as dam sites.

The problem of obtaining suitable sound equipment was examined further in the light of what was learned at Lake Mead and Chicago. A review of the geology of the area (Alcock, 1946; Bastin and Williams, 1914), particularly the nature of the sediments, was made. Also, the core data collected in the earlier investigations, made in 1935, were thoroughly studied. The conclusion reached was that sound equipment operating at a still lower frequency than was used at Chicago, perhaps 5 kilocycles, might delineate the sediments more efficiently. Shorter pulse lengths were considered desirable, but it was recognized that the complicated design problems involved in the construction of new equipment would require considerable time.

Because of the time limitation for the completion of the survey it was decided to request the Navy Department to furnish equipment similar to that employed at Chicago. Charles E. Mongan, formerly with the Marine Physics Laboratory at San Diego, was consulted. He felt that the equipment used at Chicago could be modified to produce a 6-kilocycle sound wave, but with output acoustic power reduced to 700 watts or less. The pulse lengths would still be in the range of 14–20 milliseconds. It was decided to proceed on this basis. The Navy Department generously loaned the required equipment to the Survey. Design modifications were developed by Dr. Mongan and built by the Edo Corp. (College Point, N. Y.).

The basic equipment was a later model of that used at Chicago. For the 6-kilocycle operation two new units were built: an auxiliary stack housing two auxiliary oscillators and a voltmeter, and matching networks. In the

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Plate 5.—EFFECT OF COUPLING

Tests made in Huntington Bay at site marked *TB* in Figure 15. The plate shows the bedrock for the following transducer positions: (a) on the water bottom (b) 1 foot from the bottom (c) 2 feet from the bottom (d) transducer at surface in normal operating position.

Plate 6.—EVALUATION OF SONAR PARAMETERS

Figure 1.—Comparative data on effect of pulse length on resolution in shallow-bedrock exploration of sound where less than 250 feet of overburden is present.

Figure 2.—Transducer on the water bottom; effect of receiver sensitivity.

Figure 3.—Effect of TVG, for 0- to 70-m instrument range.

Figure 4.—Pulse-length data for 1 millisecond and 5 milliseconds on the 0- to 140-m instrument range

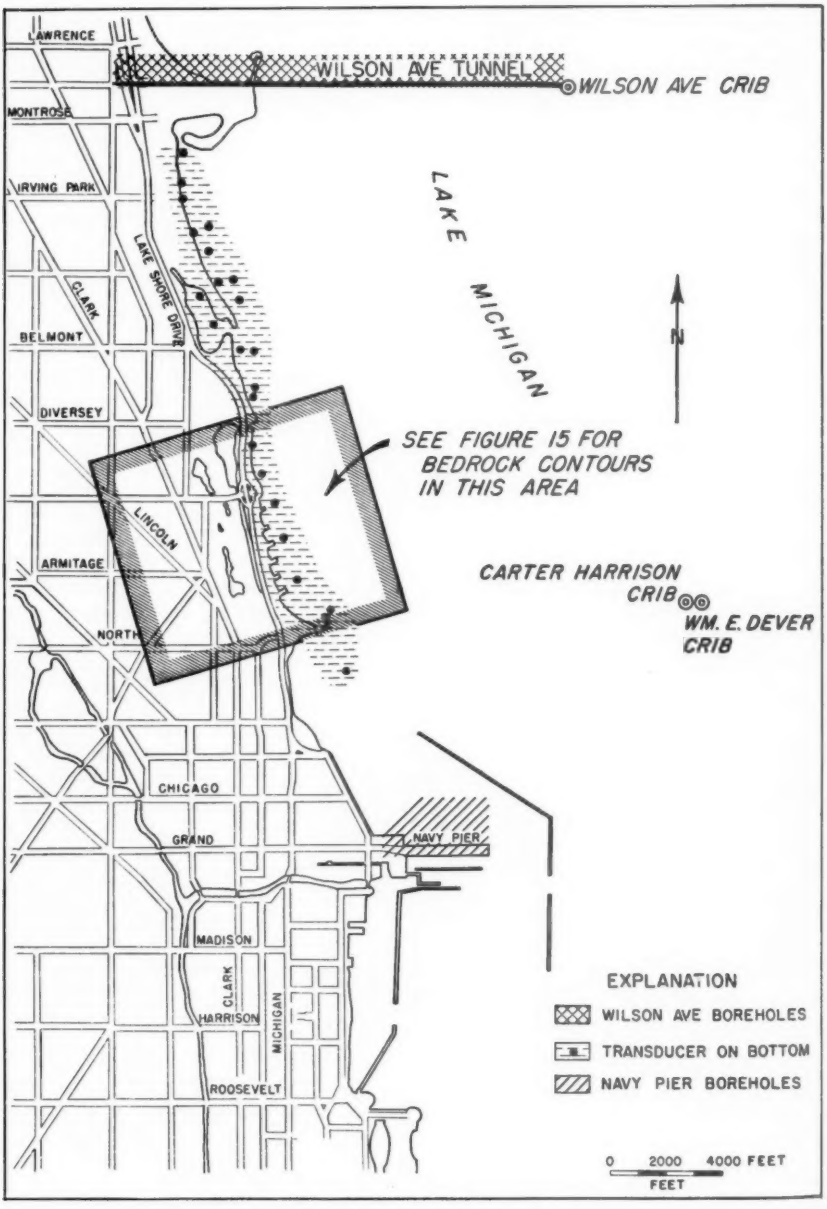

FIGURE 5.—LOCATION MAP FOR THE CHICAGO INVESTIGATION

The location of the geologic control in the lake is shown on the figure. These are the Wilson Avenue tunnel borings and Navy Pier borings. Locations of the points where the transducer was placed on the bottom also are shown.

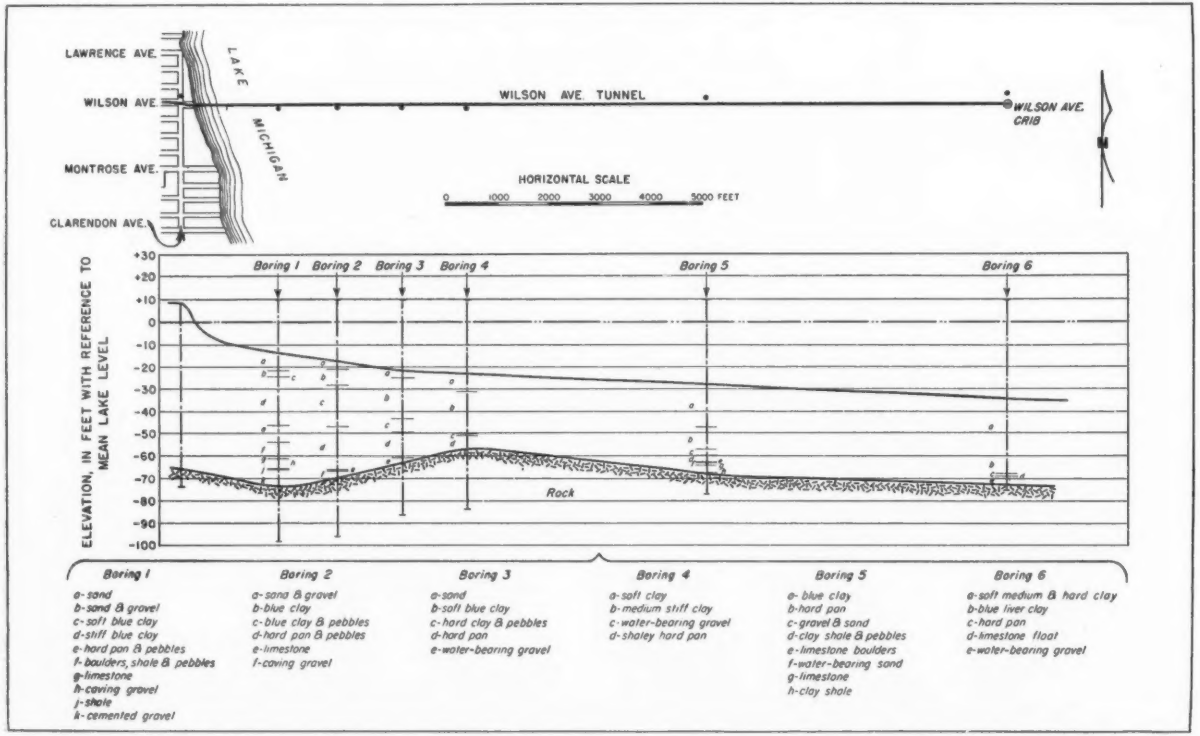

FIGURE 6.—LOCATION OF BOREHOLES ALONG WILSON AVENUE TUNNEL SYSTEM USED FOR GEOLOGIC CONTROL
The logs of these boreholes, showing textural characteristics of the overburden and the location of bedrock also are shown.

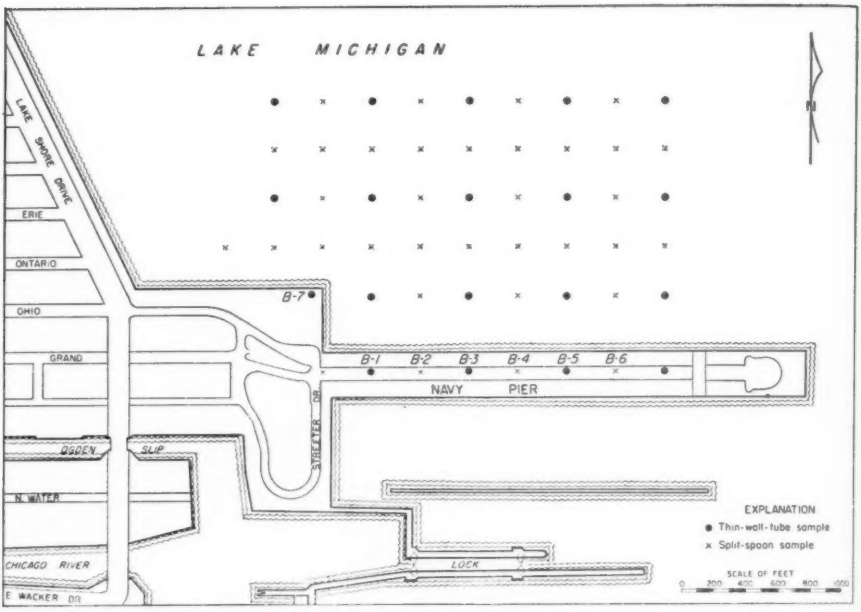

FIGURE 7.—LOCATION OF BOREHOLES IN THE NAVY PIER AREA USED FOR GEOLOGIC CONTROL

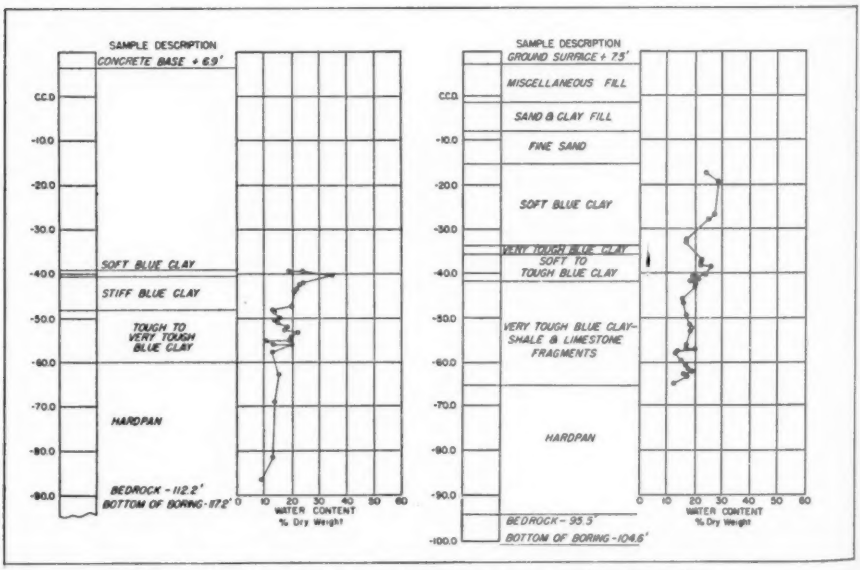

FIGURE 8.—LOGS OF TYPICAL BOREHOLES IN THE NAVY PIER AREA
Textural characteristics and water contents are shown.

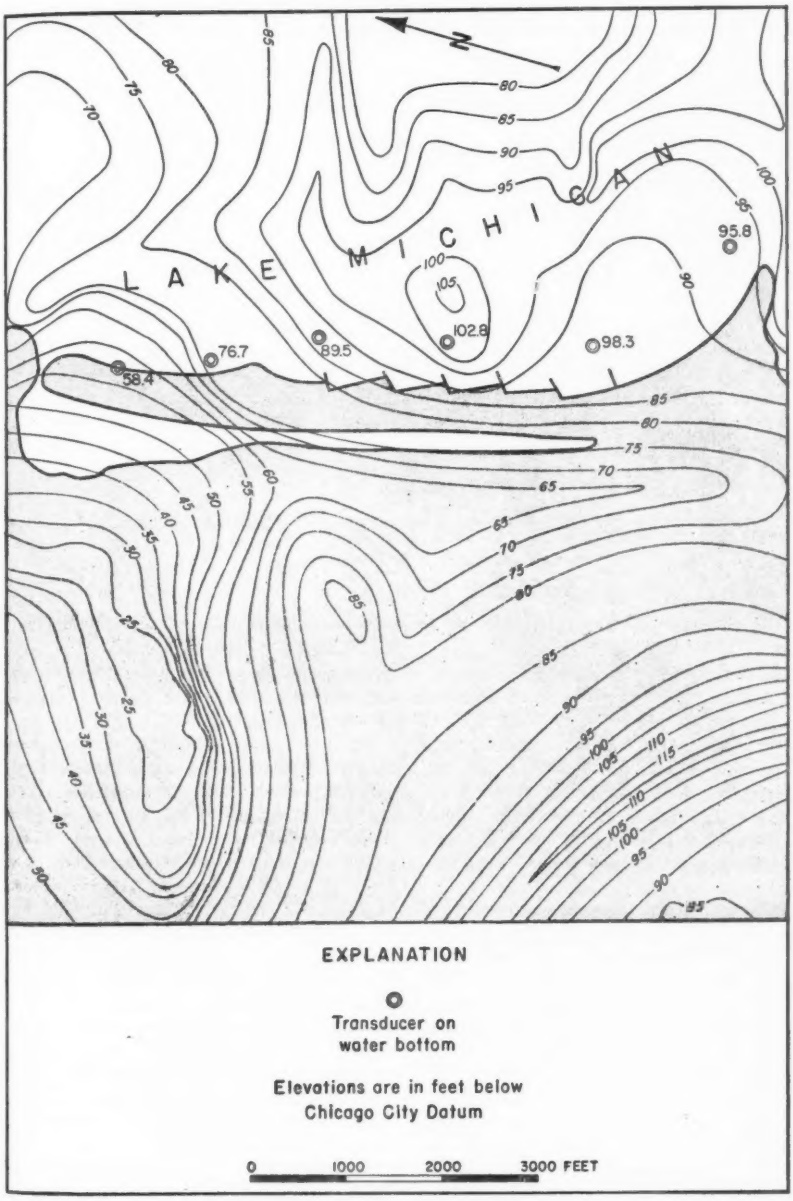

FIGURE 9.—CONTOURS ON BEDROCK IN THE VICINITY OF LINCOLN PARK
Those on shore are developed from borehole logs. Those offshore are developed from sonic data.

auxiliary stacks a large oscillator adjustable from 3 to 12 kilocycles was provided. Its output could be adjusted in amplitude by associated amplifiers and attenuators. The second auxiliary oscillator, with its associated

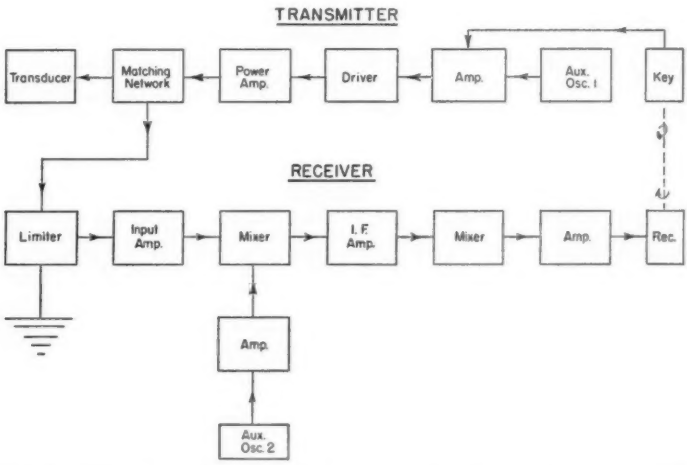

FIGURE 10.—FUNCTIONAL BLOCK DIAGRAM OF SOUNDING EQUIPMENT USED AT PASSAMAQUODDY BAY

amplifier, was used to provide a signal suitable to feed to the mixer and generate an intermediate frequency of 118 kilocycles required for processing the return signal. The matching networks provide optimum coupling between the transducer and the sonar transmitter-receiver for the various frequencies of operation. Figure 10 is a functional block diagram of the sounding equipment and its modifications.

The purpose of the Passamaquoddy investigation was to determine submarine bedrock topography and topography of the water bottom, from which could be calculated the data of greatest interest—the thickness of the intervening sediments. The study was limited to those areas where dams and certain other major engineering structures could be located. The study is described in detail in an open-file report of the Geological Survey (Smith *et al.*, 1951).

Once again the procedure followed to obtain the data was essentially that of hydrographic surveying, such as used at Lake Mead and Chicago. Eight areas shown on the location map, Figure 11, were covered. In each of these, soundings were run over range lines spaced 1000–800 feet apart and laid out to cover each area. Horizontal positioning was accomplished by the use of the standard three-point-fix procedure, using sextants. A net of shore points tied to the first-order control net of the Coast and Geodetic Survey was set up for this purpose, and all positions were thus accurately located. The details are discussed elsewhere (Smith *et al.*, 1951, p. 30).

Because sonic soundings are necessarily referred to the existing water surface, accurate vertical control was required. To ascertain specific elevations in an area of large tidal variations is a complicated problem. The tidal variations in Passamaquoddy Bay range from 13 to 28 feet and average 18.1 feet. The maximum rate of change of stage is about 1 foot in 10 minutes. The basic datum for all the results of this survey is mean low water as determined by the U. S. Coast and Geodetic Survey. The principal equipment consisted of 17 staff gages and 4 recording gages, located so that a recorder and two staff gages could be used in each of the work areas to indicate slope in any direction. The details of vertical control are to be found in the principal report (Smith *et al.*, 1951, p. 34).

An account of the geologic features of the Passamaquoddy area is to be found elsewhere (Smith *et al.*, 1951, p. 37; Upson, 1954). The rocks in the area, in the vicinity of Eastport, comprise a fairly complicated sequence of unconsolidated glacial and postglacial deposits of late Pleistocene and Recent age which rest

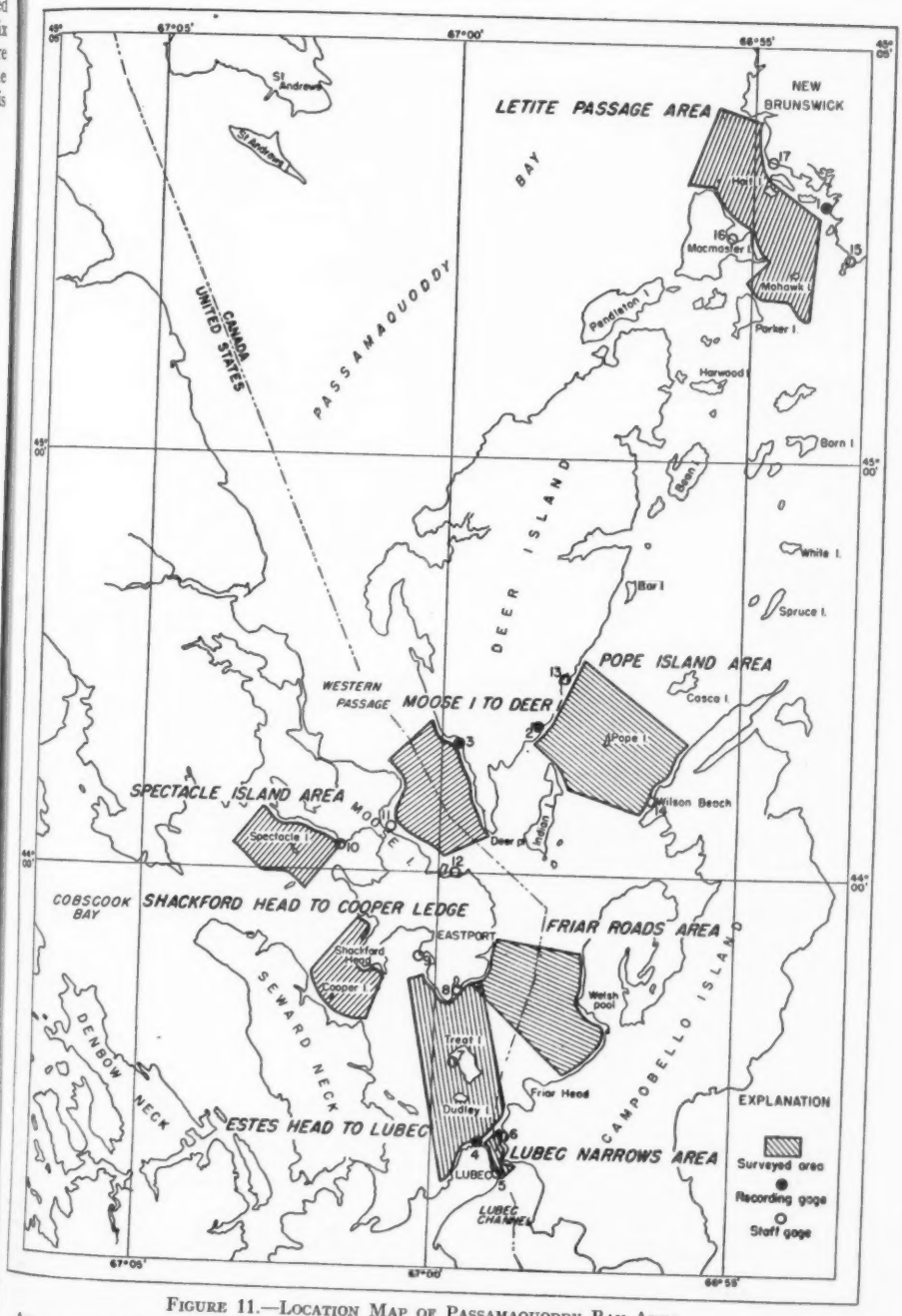

FIGURE 11.—LOCATION MAP OF PASSAMAQUODDY BAY AREA
Areas surveyed in 1951 are shown by crosshatching. Location of all water-stage gages are marked.

on an erosional surface developed on consolidated igneous, metamorphic, and sedimentary rocks of Silurian and Devonian age.

Geologic control and calculation of sound velocities were made possible by accurately methods outlined earlier. The procedure was to run directly over the lines of boring used for calibration. The area from Dudley Island to Lubec was the chief area used for this purpose. In this area the lines *A-A*, *B-B*, *C-C*,

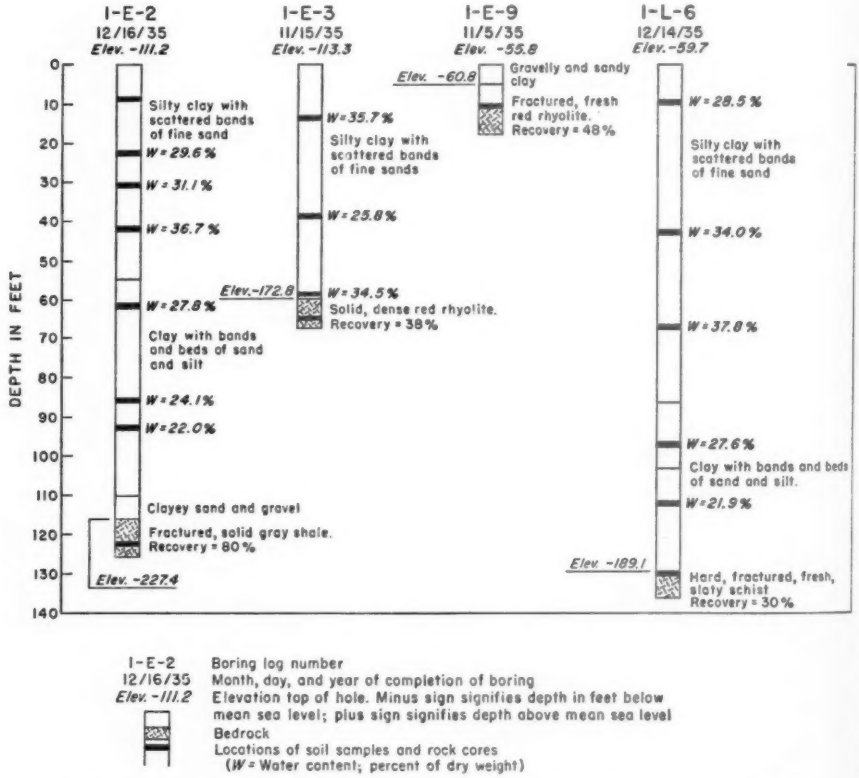
I-E-2
12/16/35
Elev. -111.2
I-E-3
11/15/35
Elev. -113.3
I-E-9
11/5/35
Elev. -55.8
I-L-6
12/14/35
Elev. -59.7
DEPTH IN FEET
ELEVATION, IN FEET BELOW M.L.W.
DEPTH, IN FEET BELOW MEAN SEA LEVEL
Silty clay with scattered bands of fine sand
W = 29.6 %
W = 31.1 %
W = 36.7 %
W = 27.8 %
Clay with bands and beds of sand and silt
W = 24.1 %
W = 22.0 %
Clayey sand and gravel
Fractured, solid gray shale. Recovery = 80 %
Elev. -227.4
W = 35.7 %
Silty clay with scattered bands of fine sands
W = 25.8 %
Elev. -172.8
W = 34.5 %
Solid, dense red rhyolite. Recovery = 38 %
Elev. -60.8
Gravelly and sandy clay
Fractured, fresh red rhyolite. Recovery = 48 %
W = 28.5 %
Silty clay with scattered bands of fine sand
W = 34.0 %
W = 37.8 %
W = 27.6 %
Clay with bands and beds of sand and silt.
W = 21.9 %
Elev. -189.1
Hard, fractured, fresh, slaty schist Recovery = 30 %
I-E-2 Boring log number
12/16/35 Month, day, and year of completion of boring
Elev. -111.2 Elevation top of hole. Minus sign signifies depth in feet below mean sea level; plus sign signifies depth above mean sea level
Bedrock
Locations of soil samples and rock cores (W = Water content; percent of dry weight)

FIGURE 12.—GRAPHIC LOGS OF TYPICAL BORINGS (1935), IN PASSAMAQUODDY BAY

logged boreholes put down during the 1935 investigations. These were in the area between Estes Head and Treat Island and that between Dudley Island and Lubec, shown in Plate 9.

Graphic logs of typical borings are shown in Figure 12. Borings I-E-2, I-E-3, and I-E-9 are in the Estes Head-Treat Island area. Their locations are shown in Plate 9. Boring I-L-6 is in the Dudley Island-Lubec area, and its location likewise is shown in Plate 9. The sediments are chiefly clay and some silt, fine sand, and gravel. The water content is a little greater than in the sediments at Chicago.

Sound velocities were calculated by the and *D-D* shown in Plate 9 were used. The average sound velocity was found to be 6250 ft./sec. The area from Estes Head to Treat Island shown in the same figure was used to check the accuracy of this value of sound velocity Each of the boring lines *A-A*, *B-B*, and *C-C* was used. A profile obtained by running over the line *B-B* is shown in Figure 13*a*. The bars represent the actual borings, and their bottoms rest on bedrock. The broken line represents the bedrock profile obtained with sonic soundings. The circle and crosses each represent a distinct run and are the points taken from the sound records to draw the bedrock profile. Figure 13*b* is a profile of the sediment struc-

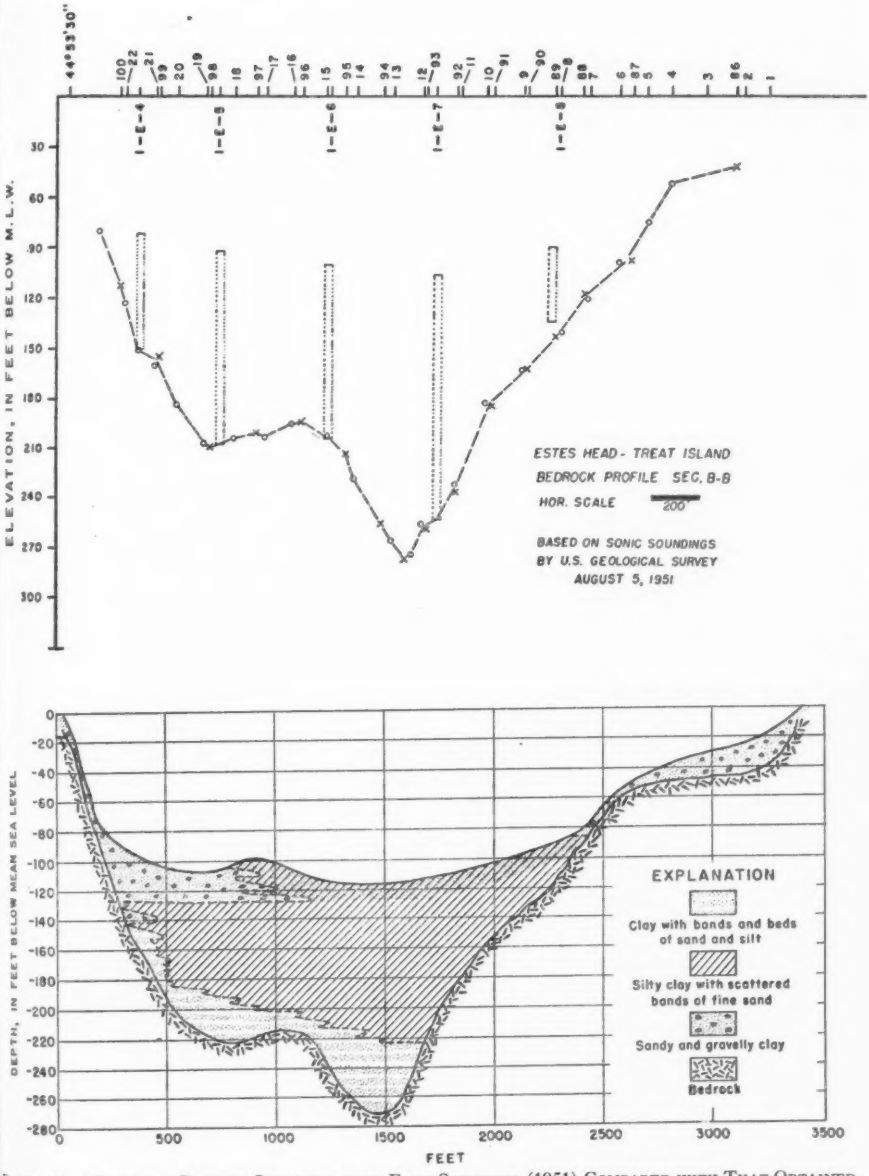
ELEVATION, IN FEET BELOW M.L.W.
ESTES HEAD - TREAT ISLAND
BEDROCK PROFILE SEC. B-B
HOR. SCALE 200'
BASED ON SONIC SOUNDINGS
BY U.S. GEOLOGICAL SURVEY
AUGUST 5, 1951
DEPTH, IN FEET BELOW MEAN SEA LEVEL
FEET
EXPLANATION
Clay with bands and beds of sand and silt
Silty clay with scattered bands of fine sand
Sandy and gravelly clay
Bedrock

FIGURE 13.—SEDIMENT PROFILE OBTAINED FROM ECHO SOUNDING (1951) COMPARED WITH THAT OBTAINED FROM CORES (1935)

Estes Head-Treat Island section *B-B*. (a) Bedrock profile from echo-sounding. Dashed line is average bedrock profile from two sets of soundings represented by crosses and circles. All cores taken in the section are shown. (b) Sediment profile based on 1935 cores.

ture in Eastport Channel along line *B-B*, for which sonic data are presented in Figure 13*a*. Plate 9 shows the submarine contours, both sediment-surface and bedrock, developed for the areas from Estes Head to Treat Island and Dudley Island to Lubec.

All sound records of the Passamaquoddy Survey were made at a frequency of 6 kilocycles. They are much clearer than those made in the earlier Chicago investigations at 11 kilocycles.

Figure 3 of Plate 3 is part of a sound record from the Pope Island-Deer Island area. It shows a contact between a clay and a sand and gravel deposit, in addition to the bedrock.

Restudy of Techniques and Redesign of Equipment

In connection with a proposed waterway survey to map bedrock and measure the sediment overburden along a Midwestern stream, a further review of the technical problems was undertaken in September 1951. The problem was to decide whether sound equipment wholly adapted to the delineation of bedrock and sediment overburden was necessary. This involved a considerable study of the technical problem as it relates to the measurement of overburden, with emphasis on the electronics necessary for efficient operation. All previously used equipment had been designed strictly for water-bottom delineation and was temporarily modified and adapted to the sediment problem. From this study it was concluded that, to be satisfactory for sediment and bedrock explorations, new equipment would have to be designed. The sponsors of the waterway survey found it necessary to postpone the project, and so the first tests of the redesigned equipment were made on Long Island Sound.

Many factors had to be considered. Critical review of earlier bedrock surveys convinced the writer and his associates that there were five essential operating conditions to be considered. These are: acoustic power, acoustic frequency, acoustic pulse length, receiver sensitivity, and finally the graphic record.

The ordinary depth sounder concerns itself principally with sufficient acoustical power to overcome propagation losses and bottom-reflection losses for the range of depth desired. Since the bottom is always present and constitutes an enormous target, the problem is relatively simple. Less than 500 watts of acoustic power can be used in practical depth sounders to measure the depth of the deepest ocean on earth.

Acoustic power of the order of 1 kilowatt or more is required for geophysical work. Although power is not the most important item in penetration, and phenomena such as reverberation have blanketing effects, it is essential to provide for power loss at each interface crossed, in addition to those in normal transmission between interfaces. The transducer faces are sufficiently large that cavitation is not a problem.

In general, power output is one of the limiting factors on depth of penetration. Normally 1 kilowatts of output power should be sufficient to secure a penetration of 1000 feet–1500 feet of sediment overburden.

The acoustic frequency generated by the equipment is probably the most important single factor in its design. The waves used in seismic work are acoustic, and though it is known that frequencies up to 15 kilocycles can be used, frequencies of 6 kilocycles or less are more satisfactory for geophysical work. It is to be expected, therefore, that acoustic frequencies adequate to penetrate sediments and to yield satisfactory information for geophysical analysis will be 6 kilocycles or less.

The operating frequency sought is therefore made as low as possible, within reason. The limit is the transducer size and beam width. The narrowest possible beam is desired to maintain a clean perpendicular crossing of each interface, thereby avoiding response from off-center energy which returns after a longer interval and extends the echo duration. Greater transducer dimensions will produce a narrower beam and direct the power downward more effectively, but cost and weight rise exponentially and thereby create a very real limit. It is possible under certain conditions to process the beam by acoustic-reflector action to narrow artificially the response without a large projector area.

High receiver sensitivity naturally contributes to the reception of weak signals, and much can be gained by its use. The receiver should be made as sensitive as the signal-to-noise ratio permits. This means that the sensi-

Plate 7.—SOUND RECORDS WITH TRANSDUCER ON THE WATER BOTTOM

(a) The effect of output power for 2-millisecond pulses on the 0- to 70-instrument range is illustrated. (b) The effect of pulse length is shown for the 0- to 70-m range.

SOUND RECORDS WITH TRANSDUCER ON THE WATER BOTTOM

COMPARISON OF PULSE-LENGTH DATA

tivity of the receiver is related to the intensity of the useful signal and the intensity of extraneous noises. Useful gain obtains in having a receiver of sensitivity at least as high as the level of the useful signal in the presence of disturbing noise. In general it has been found that a 10-microvolt sensitivity at the receiver is satisfactory. The importance of a high receiver sensitivity cannot be overemphasized. Many of the returning echoes are very weak and would be lost without a sufficiently sensitive receiver. Deep echoes often occur after reverberation has subsided, when the extraneous noise level is low and a weak signal can be "read" satisfactorily.

During the work at Passamaquoddy Bay, good penetration was secured because the output power and 6-kilocycle frequency were satisfactory. Usually only the bedrock trace was discernible, however, although in some places a sediment interface appeared. Long pulse lengths with consequent overlapping of traces were probably responsible.

Owing to the limitations of the available instruments, the pulse lengths employed at Passamaquoddy were 14–20 milliseconds. As a result, echoes from the more closely spaced sediment strata were obscured because of overlapping. The length of a transmitted pulse is the time duration of a single transmission of sonic energy. In order to obtain depth penetration, the pulse must have an appropriate duration in addition to magnitude. A long pulse produces a long echo which may obscure the area of the chart where other traces are being imprinted. It is therefore advisable to obtain as short an echo as possible, consistent with penetration. The instrument should be designed to produce pulse lengths variable from 1 to 9 milliseconds. Exceptionally good results should be obtained with 1- and 2-millisecond pulses in shallow sediments. Good separation of echoes from interfaces as close together as 10 feet ought then to be possible. Examples of these will be given in connection with the discussion of results on Long Island Sound.

During the review of the technical problem, the design of the basic type of presentation was considered in great detail. The recorder used during the Passamaquoddy investigation was simply a normal depth recorder, designed primarily to reproduce the water bottom only. Consequently, it printed either dark or not at all. Thus some of the weaker echoes were probably lost.

It is conceivable that even a fair bedrock echo could be lost when the normal depth recorder is used. The need for a recorder with greatly enhanced resolution is apparent. It is necessary to discriminate between signals of nearly the same magnitude. Technically, this property of distinguishing between nearly equal signals is measured by the so-called dynamic range—that is, the number of shades of gray that can be resolved.

This end can be achieved principally by use of a paper and a printing method that produces a degree of darkness of trace that is really a linear function of current strength. In the recorder finally chosen about 18 different shades of grayness can be distinguished, in contrast to the normal depth recorder which prints either black or not at all. The recorder is essentially a facsimile recorder modified for use in sonar work. It provides a dynamic range of 20 decibels compared to 6 decibels for the normal depth recorder. The recorder uses a chemically treated paper which must be handled in a humidor. There are adjustments on the recorder which facilitate the setting of the intensity of the weakest and the strongest signal to-marking ratio—that is, the white and black level.

When normal depth recorders are used for sediment delineation, the upper part of the chart is generally blackened when the gain is turned to levels that will bring in the feebler depth traces, and much of the record is ordinarily lost. This defect is corrected by inserting a time circuit which controls the gain as a function of time, usually referred to as TVG. This means that after the pulse leaves the transducer, when the water is fully ensonified and reverberation is high, the receiver is automatically desensitized in accordance with a preestablished time curve. This process prevents large black chart areas at the high-echo-energy depths, and low-energy sub-bottom traces are received more distinctly.

The general specifications and overall design work were done by the author in consultation with George B. Cummings, (formerly Chief Civilian Engineer, Sonar Design Branch, Bureau of Ships; now assistant to President,

PLATE 8.—COMPARISON OF PULSE-LENGTH DATA

For the instrument ranges 0 to 140 m and 0 to 420 m

Telephonics Corp., Huntington, N. Y.). Design work was completed in 1952 and the instrument built early in 1953. A functional block diagram is shown in Figure 14. Acoustic frequencies of 6, 11.5, and 16 kilocycles are provided. Acoustic powers of 250 watts to 2500 watts, adjustable within transducer limitations, are provided. Output electrical power varies over the range 500 watts–5 kilowatts. Three depth ranges, each having a limited range of pulse length as follows, are provided.

	(a)	(b)	(c)
Depth range (meters)	0–70	0–140	0–420
Pulse length (variable) (ms)	1–3	1–5	1–9

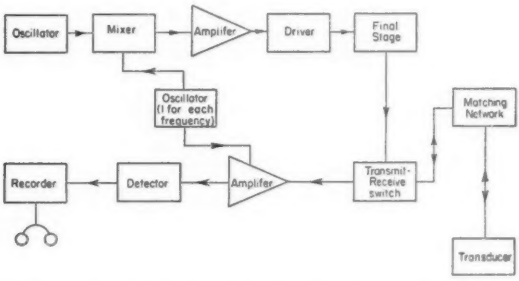

FIGURE 14.—FUNCTIONAL BLOCK DIAGRAM OF LONG ISLAND EQUIPMENT

The present transducer has a natural resonance at 11 kilocycles. It is driven-off resonance, as with the Passamaquoddy equipment designed by Dr. Mongan, when used for 6-kilocycle exploration. The output power is then down about 10 decibels from that of the normal 11-kilocycle operation. It is possible to modify the present equipment for operation at 3 kilocycles and still have adequate power for sediment penetration. as there is probably less attenuation at 3 kilocycles than at 6 kilocycles during penetration of sediments.

LONG ISLAND SOUND INVESTIGATION

A series of tests was run in Long Island Sound to evaluate the performance of the new equipment and its suitability for bedrock and sediment investigations. Preliminary tests were made to establish the optimum acoustic frequency and pulse length. First a check on frequency response of sediments was made using a pulse length of 2 milliseconds and frequencies of 6 kilocycles and 11 kilocycles. The test site was in the western end of Long Isla Sound, north of Manhasset Bay. It is on t range line *AA*, near fix 10, shown on the loc tion map, Figure 15. Figure 1*a* of Plate 1 sho the results of a test run at 11 kilocycles, a Figure 1*b* (Pl. 1) shows a run over the sam course in the opposite direction at 6 kilocycl The bedrock topography is clearly delineated the run made at 6 kilocycles; in that made at kilocycles it is definite but not nearly as cle In the 6-kilocycle test details of the overburd are clear; they show also in the 11-kilocyc test but are hazy.

Two multiple bedrock echoes are shown i the 6-kilocycle test and illustrate the phenom ena discussed under multiple sediment echos The first, labeled "Bedrock 2d echo" in Figu 1*b* (Pl. 1), is a reflection of a part of t original bedrock echo, labeled "Top of Be rock" in the figure, which has been reflected a the water surface back to the water botto where it was reflected once more and returne to the transducer. The other bedrock multip echo, labeled "Bedrock 3d Echo" in Figure 1 (Pl. 1), shown below that just described, is part of the original bedrock echo which ha undergone a reflection at the water surface passed through the water bottom and onc more been reflected from the bedrock bac through the water bottom to the transducer The conclusions are based on path measur ments on the sound record, correction bei made for the difference in sound velocities sediment and water. Only one multiple bedroc echo is observed in the 11-kilocycle test. corresponds to a part of the return echo fro the bedrock which has undergone reflecti from the water surface to the water botto where it was returned to the transducer. N second reflection from the bedrock is observe as in the 6-kilocycle test. This probably ind

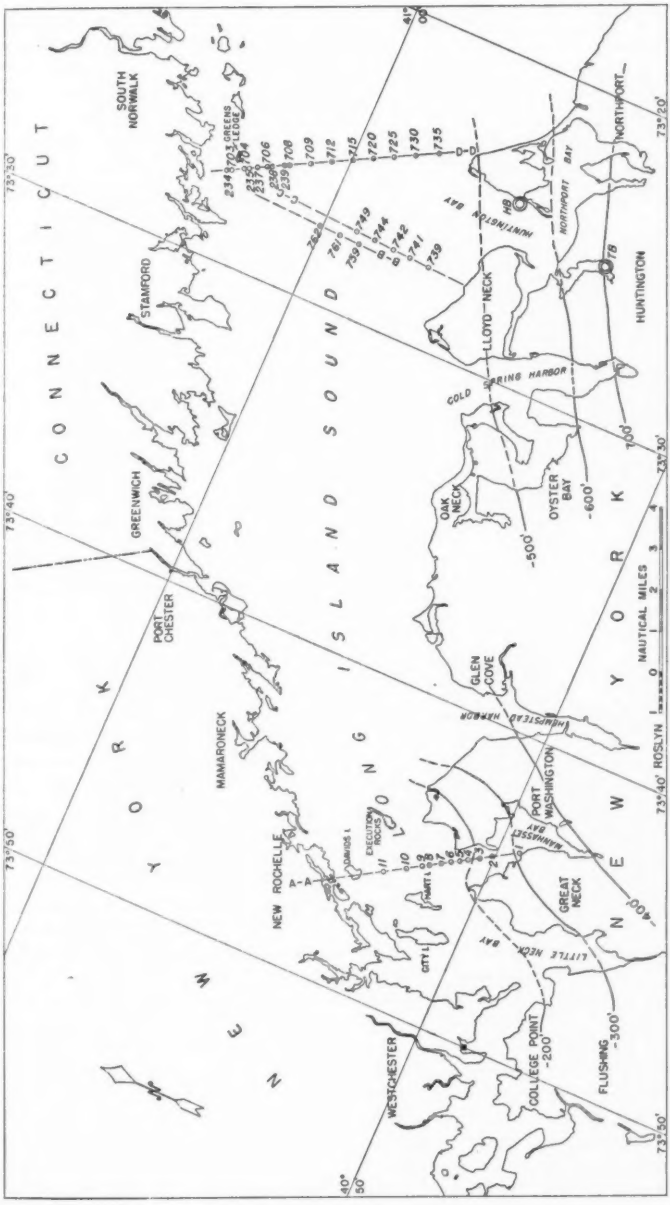

FIGURE 15.—LOCATION MAP OF THE LONG ISLAND SOUND INVESTIGATION

Test ranges are shown together with contours (based on well logs) on bedrock in the vicinity of each.

cates a much greater attentuation in the sediments for an 11-kilocycle sound wave than for 6 kilocycles. The 11-kilocycle trace is similar to that observed in the Chicago investigation, except for the better resolution afforded by the bedrock. Travel times to bedrock and return are known from the sound records. Interpolation of contours was used to find depths to bedrock at each of the chosen fixes. Fixes 1 and 4 are observed to be close to known contours.

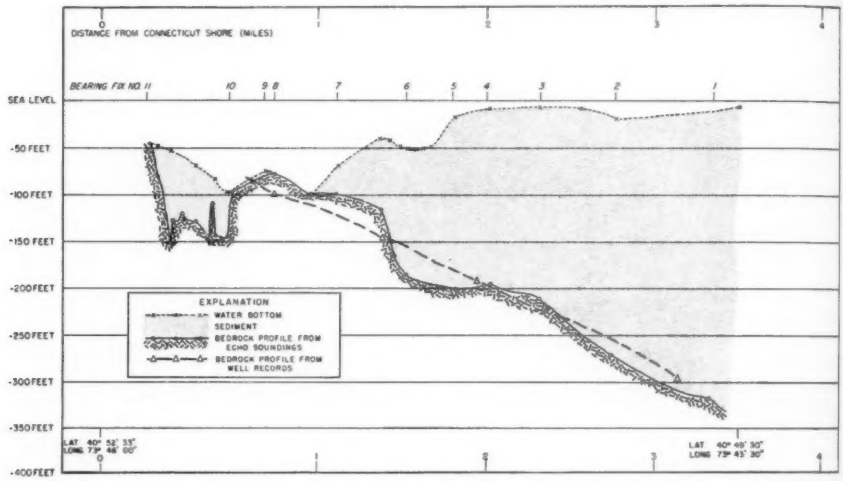

FIGURE 16.—PROFILE ACROSS LONG ISLAND SOUND, JUST NORTH OF MANHASSET BAY
It is along the range line *AA* of Figure 15.

2-millisecond pulses and a more sensitive recorder.

Preliminary tests were also run for the effect of pulse length. A typical sound record is shown in Figure 1 (Pl. 6). The effect on resolution is shown for pulses of 1, 2, 4, 5, and 9 milliseconds at 6 kilocycles, using the 0- to 70-m range of the equipment. A 1-millisecond pulse gives the sharpest record, but it is less intense than that obtained using 2 milliseconds. The longer pulses produce a trace that is not sharp and that overlaps other echoes. For example, the water-bottom echo obliterated the bedrock echo in the test when 9-millisecond pulses were used. It appears that 2-millisecond pulses are better suited for the range of 0–70 m. This question will be discussed more fully later.

Sound velocities were established from data taken on range *AA*, shown in the western part of the location map, Figure 15. A location near fix 3 on range *AA* was used for this purpose. The contours on bedrock shown in this part of the map—that is, those which cross Manhasset Bay—are based on logs of wells on Long Island (Suter *et al.*, 1949, Pl. 8; Fuller, 1914). It is assumed that they represent correct depths to The value of sound velocity found from this procedure was 5400 ft./sec.

A second set of tests was run to ascertain whether a bedrock profile across Long Island Sound could be obtained from acoustic exploration. For this purpose range line *AA* was established, as shown on Figure 15, from a point in Manhasset Bay, marked fix 1, to a point near Davids Island just north of fix 11. Good penetration was obtained for the most part, although it was poor at a few locations. Profiles of the bedrock and of the sediment surface are shown in Figure 16 for the line *AA*. Both were obtained by sonic methods. The heavy dashed line in the same figure shows the corresponding profile on bedrock obtained from well records. The points used to plot the profile, shown by small triangles, are the intersections of the bedrock contours drawn from well records with range line *AA*. Figure 2 (Pl. 1) shows the photographs of original sound records near fix 3, range *AA*, of Figure 15. An attempt has been made to identify the geologic formations. The sound traces tentatively identified in Figure 2 (Pl. 1) as the top of the Raritan Formation, the top of the Lloyd Sand Member of that for-

mation, and the top of the bedrock check reasonably well with the expected positions of those surfaces as shown by Suter *et al.* (1949, Pls. 14, 11, 8, respectively). It is possible, and even probable, however, that the Cretaceous sediments were removed by preglacial or interglacial erosion and that the traces tentatively identified as the upper surface of the Raritan and the Lloyd actually represent younger deposits, and this possibility must be kept in mind. In Figure 2*a* (Pl. 1) the bedrock echo is weak in the photograph but distinct on the original sound record. The overburden appears well developed. Figure 2*b* (Pl. 1) shows another part of the sound record from the same area. Here the bedrock is clear, but the record for the overburden just above bedrock is not well developed. This example illustrates some of the difficulties that have not been resolved. Figure 1 (Pl. 1) shows a sound record between fixes 10 and 11 on range *AA*. South of fix 10 the sound records are similar to those shown in Figure 2 (Pl. 1). There were a few locations at which penetration was poor. It was not ascertained whether the effect was caused by inadequate output power or low level of receiver sensitivity, or whether it was due to an inherently poor sediment response.

To extend the data observed on range *AA*, three additional lines, *BB*, *CC*, and *DD*, were run. They are shown on Figure 15. The ranges begin at points north of Huntington Bay and are directed toward Greens Ledge just off the Connecticut shore. Only the line *DD* was run completely across the sound. A profile, with vertical scale considerably exaggerated, of range *DD* is shown in Figure 17. Both bedrock and top of sediment are shown. All data are from sound records. Another, similar profile for range *BB* is shown in Figure 18. Whether the deep gorge between fixes 741 and 744 is related to the gorge shown between fixes 715 and 730 on Figure 17 cannot be determined from the available data.

Plate 2 shows a sound record on range *DD*. The location is shown by line *AA'* at the top left of Figure 17. Fixes 033 to 037 are included. Considerable detail is exhibited in the overburden. This record was taken with the 0- to 70-m range of the instrument. Figure 1 (Pl. 4) is another part of the sound record, taken on the 0- to 140-m range in the vicinity of fixes 708 and 709. Fix 709 is to the left just off the figure. No attempt has been made to identify the geologic formations.

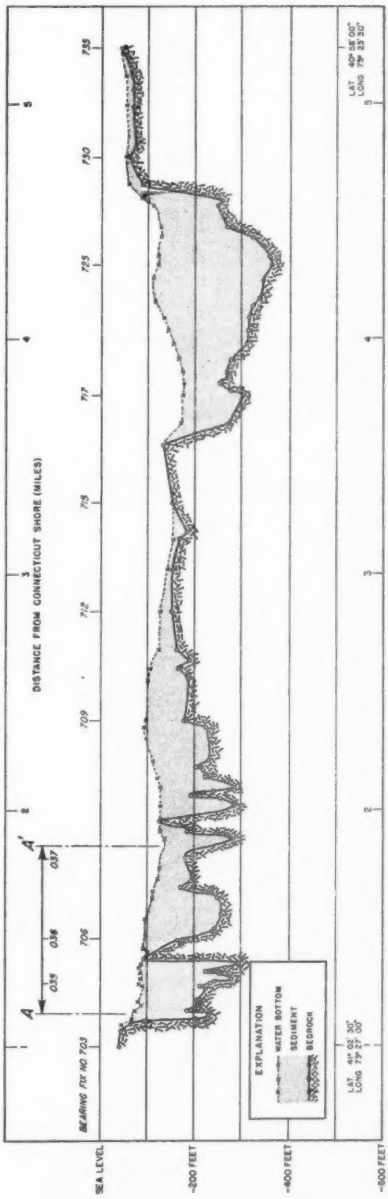

FIGURE 17.—PROFILES OF BEDROCK AND WATER BOTTOM ACROSS LONG ISLAND SOUND JUST NORTH OF HUNTINGTON BAY
The line is range *DD* of Figure 15.

Tests with the Transducer Directly on the Water Bottom

Three tests were run to ascertain the influence of coupling on depth of penetration. Plate 5 shows the sound record for this experiment. It was run in the south end of Huntington Bay in about 15 feet of water. The location is marked *TB* on the location map, Figure 15. Frequency used was 6 kilocycles with 5-millisecond pulses at maximum range of the equipment, 0–420 m. Power level was kept constant at near peak, and full receiver sensitivity was used.

In the first test the transducer was lowered to the bottom. That is, it was placed on top of the sediments so that almost no bulk water lay between the diaphragm and the sediments. The transducer was in the same position throughout the test. On the right of Plate 5 the section marked "Transducer on water bottom" shows the results with that position. The right third of the picture corresponds to an adjustment of recorder gray level and the part to the left of this is the record for good adjustment. About 750–800 feet of penetration is shown. The Cretaceous formations are shown because they are present beneath the adjacent land and can reasonably be inferred to be present beneath Huntington Bay. The approximate contacts shown correlate reasonably well with the positions expected from extrapolation of well data. The top of the Lloyd Sand Member of the Raritan Formation is not as clearly defined as is bedrock at the bottom of the Lloyd.

For the second test the transducer was raised from the bottom so that 1 foot of water separated the sediment top from the transducer diaphragm, as indicated on the record. Note that the stratification is still fairly well defined, although bedrock appears somewhat hazy. The transducer was next raised to 2 feet off the bottom. Now the stratification is still less distinct and bedrock increasingly hazy.

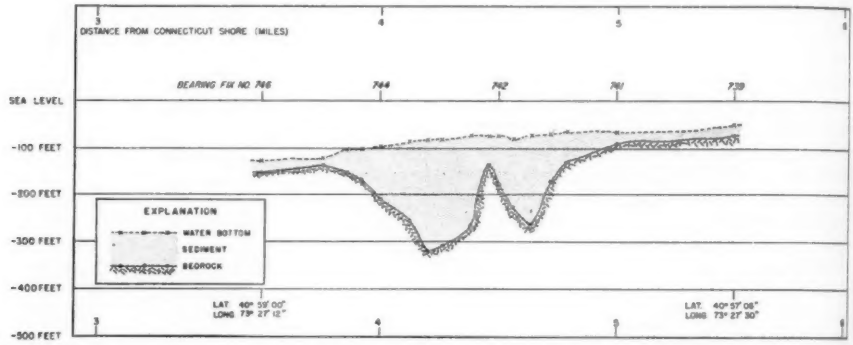

FIGURE 18.—Profiles of Bedrock and Water Bottom of a Part of Long Island Sound just West of Range *DD* Shown in Figure 17

It is range *BB* of Figure 15.

For test three, the transducer was raised to its normal operating position a few feet below the surface. Bedrock can still be identified, but the boundaries of the formations and the bedrock are not nearly so distinct. No tests were made at reduced power, shorter range, or shorter pulse length.

To find the depth of the firm trace marked Bedrock the procedure is as follows: This trace appears 3.75 inches from the water bottom. The recorder stylus travels over 1 inch of paper in 0.00711 second. For 3.75 inches the time consumed is 3.75 (0.00711) = 0.0267 second. This is the total time required for the sound to travel from the transducer to suspected bedrock and back to transducer. In the absence of definite acoustic velocities, seismic values determined by Ewing (1950, p. 889) and by Oliver and Drake (1951, p. 1293) were used as a check. For unconsolidated sediments, the seismic velocity determined at Cape May, New Jersey, was 5400 ft./sec. For Ewing's New York profile the value was 5700 ft./sec. Oliver and Drake give a value of 5725 ft./sec. at SS21, which is close to Huntington Bay. With these values the distance to bedrock at Huntington Bay is found to be 760 and 800 feet, respectively, as compared with about 725 feet determined by extrapolation from existing well records (Suter *et al.*, 1949, Pl. 8). The value of 5400 ft./sec. for

acoustic velocity corresponds to that found earlier from well records in Manhasset Bay.

These tests do show that proper coupling of the transducer to the sediments is vital and that considerable loss of power occurs when the transducer is inefficiently coupled. No borings to bedrock are available in the immediate vicinity, and no further attempt will be made to arrive at a closer estimate of the depth to bedrock.

Evaluation of Sonar Parameters

Additional tests were run for the most part at shorter ranges to evaluate the effects of receiver sensitivity, power level, pulse length, black level, and TVG. These tests were made in the southern part of Huntington Bay at the site marked *HB* on Figure 15.

Figure 2 (Pl. 6) is a record made to test receiver sensitivity at a frequency of 6 kilocycles, with 1-millisecond pulses on the 0- to 70-m range. Power level is 100 per cent. The materials penetrated are unconsolidated sediments. On the left receiver sensitivity is 100 per cent; on the right it is 75 per cent. The difference in record clarity is striking. At the lower level of receiver sensitivity much of the record is lost, especially in the deeper rock. The dark band near the upper end of the chart is caused by reverberation. Reverberation is reduced by the TVG circuit. At 75-per cent receiver sensitivity, the automatically obtained reverberation control is overridden earlier and allows the dark reverberation trace to appear sooner.

Effects of output power level are shown in Figure 2 (Pl. 4) and Plate 7. On the left of Figure 2 (Pl. 4), data taken at 6 kilocycles with 3-millisecond pulses are shown for the 0- to 140-m range. On the extreme left are data taken at 75-per cent output power level, and adjacent are data taken at 100-per cent output power level. The loss in detail at the lower power level is apparent. The increased power, however, has the disadvantage of blanking out, by reverberation, the response of sediments lying near the water bottom.

Two solutions to this dilemma are feasible. The difficulty can be met by making two runs over the same rock, one at high power to delineate the deeper rocks, and one at reduced power to develop the shallower. Or, on a single run, a time-varied gain of the receiver (TVG) can be used at nearly full power level.

Effects of output power level are illustrated on the right half of Plate 7, which shows data run at 6 kilocycles, on the 0- to 70-m range of the instrument with 2-millisecond pulses. At 75-per cent power level the sedimentary layers near the water bottom are distinct, but those below 130 feet are weak or missing. At 100-per cent power level the layers near the water bottom are obscured, but those below 130 feet are quite distinct.

TVG has the effect of automatically introducing full receiver gain when sound from the deeper rocks is recorded and again automatically reducing the receiver sensitivity when reception from the shallower rocks is recorded. The reduced receiver gain eliminates the effect of feebler echoes and reduces the intensity of stronger ones, which is equivalent to reducing output power for shallow reception and increasing it for deep conditions. The right half of Figure 2 (Pl. 4) shows the effect of TVG for 1-millisecond pulses. Full TVG increases the intensity of the record for deep echoes and increases resolution for strata from 100 to 300 feet below water bottom. The record shown in the figure is not the best adjustment for the conditions but nevertheless shows the possibilities of accurately controlled TVG. Figure 3 (Pl. 6) illustrates the effect of TVG for 3-millisecond pulses and shows much better results than those shown in Figure 2 (Pl. 4).

The effect of pulse length is illustrated in Plate 7 and Figure 4 (Pl. 6). The left half of Plate 7 shows data taken on the 0- to 70-meter range of the instrument, at 6 kilocycles with 100-per cent power level and constant receiver sensitivity. The data are for pulse lengths of 1, 2, and 3 milliseconds. The 3-millisecond records are not as clear as the others. The 1-millisecond record shows much clearly resolved detail in the sediments down to 200 feet below the water bottom. The 2-millisecond record shows good resolution in the depth range 130–300 feet. Records in the range 0–130 feet made with 2-millisecond pulses are good but not quite so clear as for 1-millisecond pulses.

Figure 4 (Pl. 6) shows pulse-length data taken for the 0- to 140-m instrument range, at a frequency of 6 kilocycles with 100-per cent receiver sensitivity and nearly full power level. The records for pulse lengths greater than 1 millisecond are good up to a maximum distance of 210 feet below the water bottom. Below this they are lost. The vagueness of the record where the figure is marked "vessel rolling" is probably caused by a spurious noise effect resulting from the roll of the ship. When 5-millisecond pulses

are used the entire record is usable up to about 600 feet below the water bottom. The longer pulses do not, however, give as good resolution in shallower layers as the shorter ones. The left half of Plate 8 shows data similar to those of Plate 7, for pulse lengths of 1 millisecond and 3 milliseconds on the 0- to 140-m range of the instrument. The 3-millisecond pulse appears to give a more satisfactory record up to 560 feet than does a 1-millisecond pulse. The 1-millisecond records are not as intense as those for 3 milliseconds. The right half of Plate 8 shows data taken on the 0- to 420-m instrument range for a frequency of 6 kilocycles; the receiver sensitivity is 100 per cent and output power constant is but a little less than full level. Data are shown for 1-millisecond, 5-millisecond, and 9-millisecond pulse lengths. The longer pulse lengths show much better penetration.

The last heavy trace in Plate 8 at about 650 feet is probably bedrock. The traces below are most likely multiple echoes from within the sediments, but the possibility of effects from acoustic discontinuities below the top of the bedrock have not been eliminated.

The results of the Long Island tests show that low-frequency sound is satisfactory for shallow bedrock and sediment exploration under water. As is usual in geophyscial exploration, adequate geologic reconnaissance and control established by test drilling are necessary. In this work the lowest frequency used, 6 kilocycles, gave the best results. It is possible, however, that a still lower frequency, for example, 3 kilocycles, would prove more efficient in the delineation of sediments.

The effect of pulse length is important. A 2-millisecond pulse produces better results in the shallower sediments, whereas longer pulses of 5–9 milliseconds are better for those lying deeper. This is because a greater amount of energy is present in the longer pulses, and is thus available not only for overcoming acoustic attenuation in sediments traversed during propagation, but for providing adequate energy for penetration after reflection losses at numerous sediment interfaces.

The output power level of the transducer, receiver sensitivity, and dynamic range of the recorder also are important factors.

The considerably increased performance obtained by placing the transducer directly on the bottom illustrates the effect of coupling. Little is known about this. Much research is needed to determine the acoustic response of different kinds of sediments. In particular, data are needed on the attenuation coefficients of sound and also on the velocity of sound while penetrating different types of sediments.

The effect of water content and rock density is important. Data on texture also are needed for sand, silt, and clay of various kinds. Data on the reflection coefficients of various sediment interfaces are important. They form a base for calculating the transmitted and reflected power at such interfaces and should facilitate predictions on the depth of penetration that may be had in a given formation. (Reflection coefficients can probably be calculated from sound velocities and rock densities).

A knowledge of proper coupling procedures is necessary to get maximum power into the sediments. If such were known it probably would not be difficult to apply acoustic techniques to land exploration, at least in those cases where the underlying sediments are partially filled with water.

In spite of the meagerness of existing knowledge of the acoustic response of sediments, low-frequency acoustic techniques have been found to be satisfactory for shallow underwater exploration. Much has been accomplished in the way of instrument design.

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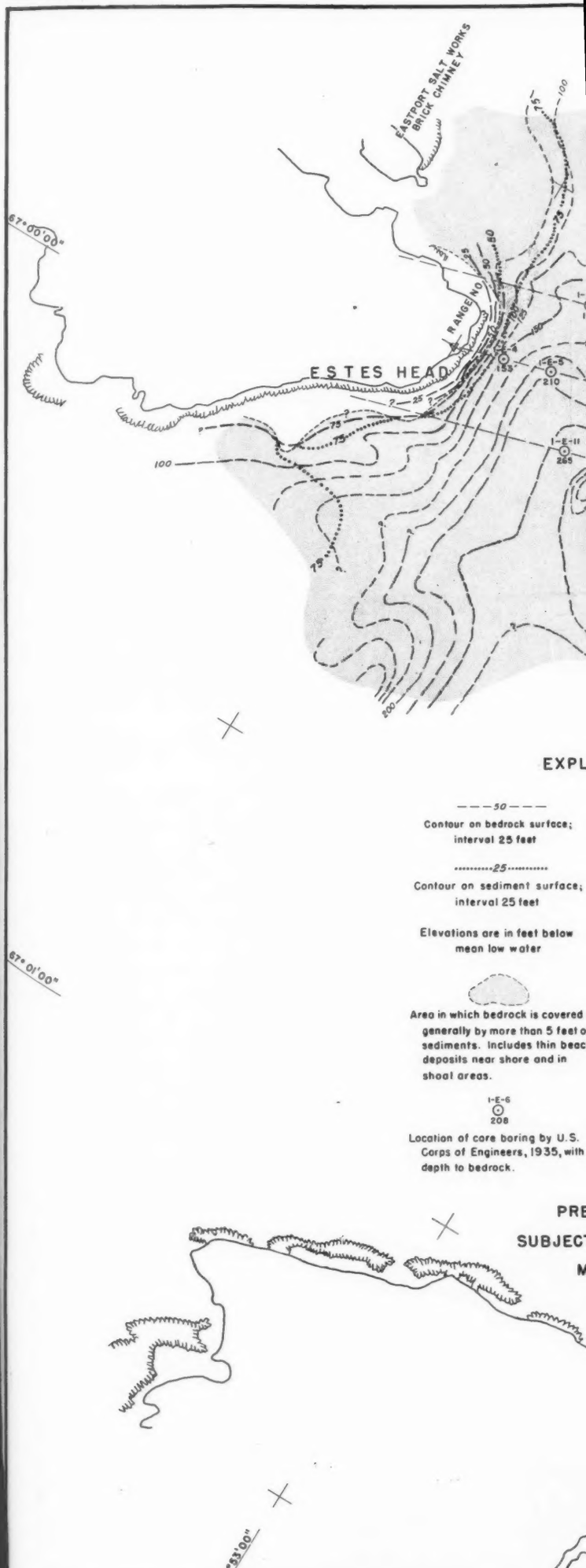

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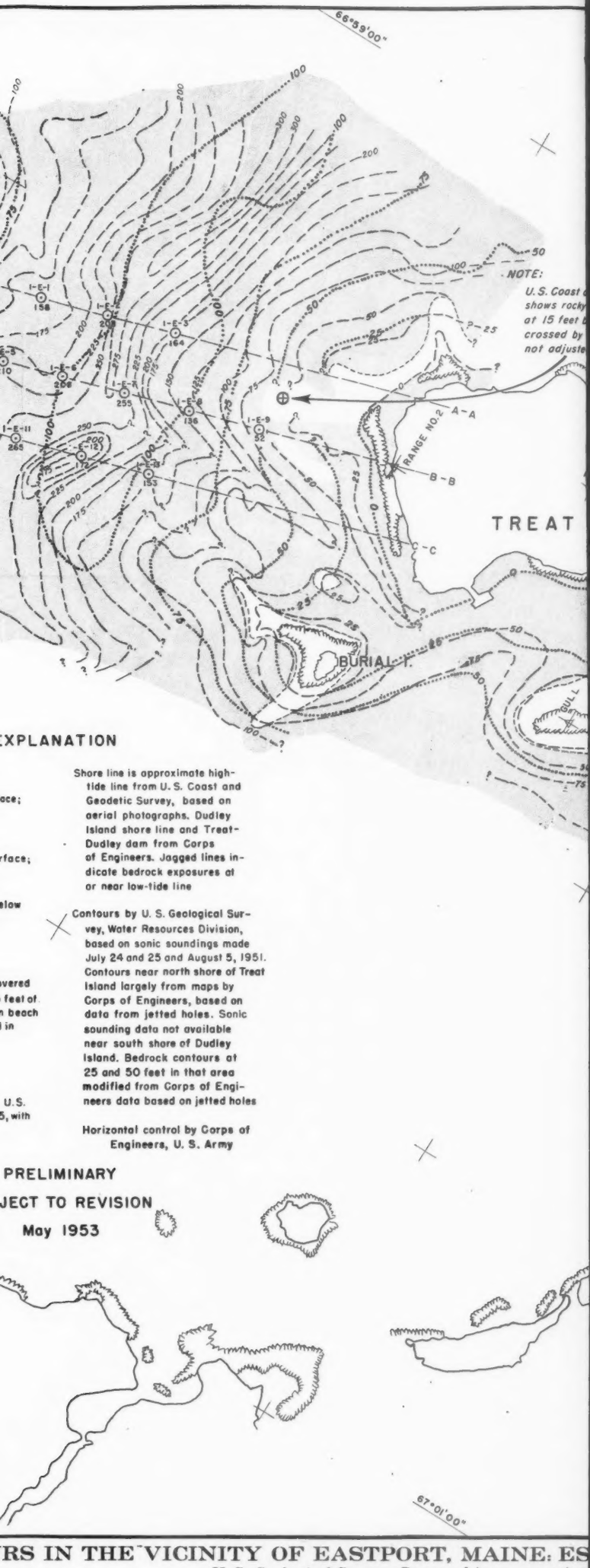

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Shore line is approximate high-tide line from U. S. Coast and Geodetic Survey, based on aerial photographs. Dudley Island shore line and Treat-Dudley dam from Corps of Engineers. Jagged lines indicate bedrock exposures at or near low-tide line

Contours by U. S. Geological Survey, Water Resources Division, based on sonic soundings made July 24 and 25 and August 5, 1951. Contours near north shore of Treat Island largely from maps by Corps of Engineers, based on data from jetted holes. Sonic sounding data not available near south shore of Dudley Island. Bedrock contours at 25 and 50 feet in that area modified from Corps of Engineers data based on jetted holes

Horizontal control by Corps of Engineers, U. S. Army

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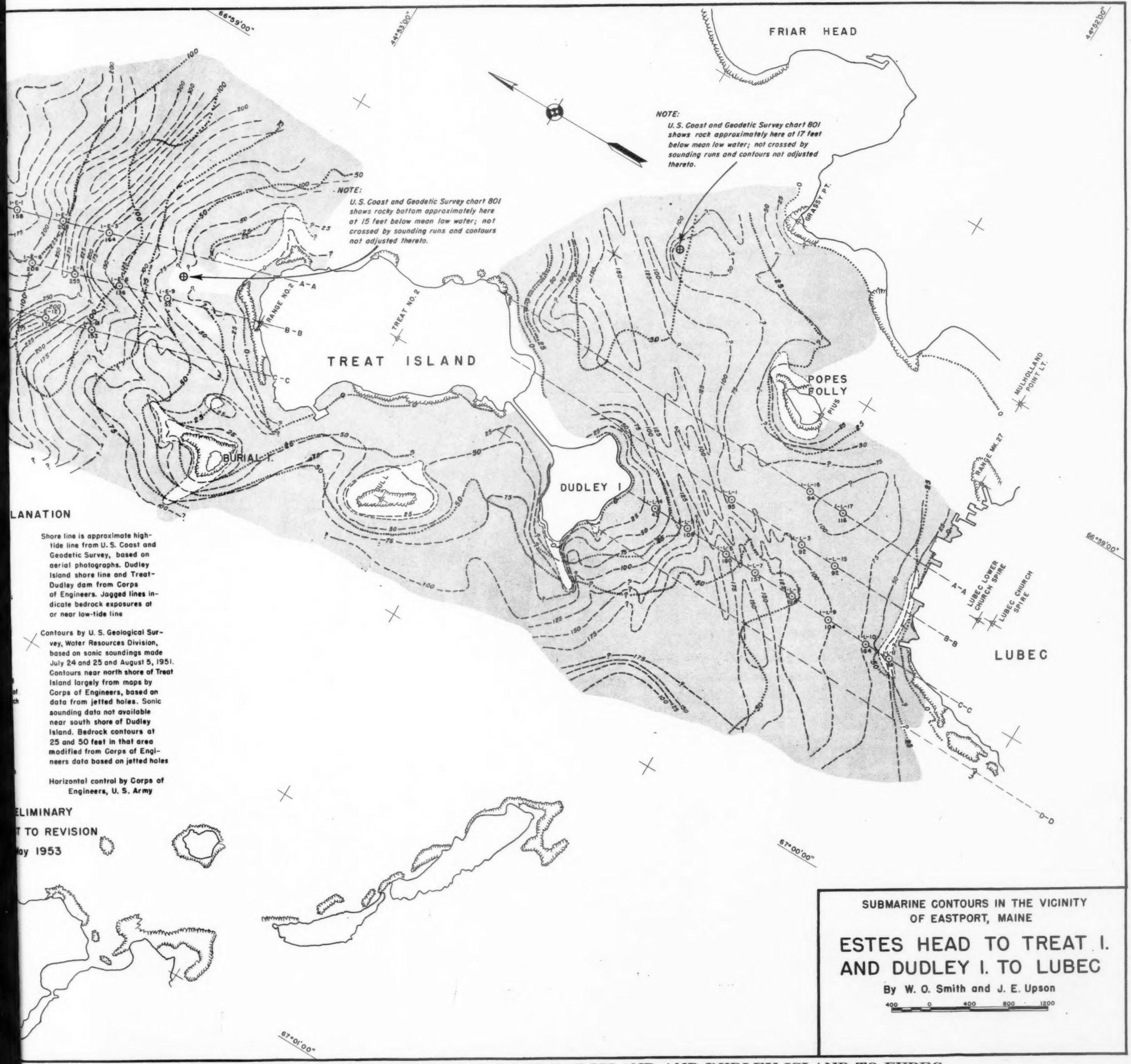
U. S. Geological Survey, Prepared in co-operation



LANATION

Shore line is approximate high-tide line from U. S. Coast and Geodetic Survey, based on aerial photographs. Dudley Island shore line and Treat-Dudley dam from Corps of Engineers. Jagged lines indicate bedrock exposures at or near low-tide line

Contours by U. S. Geological Survey, Water Resources Division, based on sonic soundings made July 24 and 25 and August 5, 1951. Contours near north shore of Treat Island largely from maps by Corps of Engineers, based on data from jetted holes. Sonic sounding data not available near south shore of Dudley Island. Bedrock contours at 25 and 50 feet in that area modified from Corps of Engineers data based on jetted holes

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SUBMARINE CONTOURS IN THE VICINITY OF EASTPORT, MAINE

ESTES HEAD TO TREAT I. AND DUDLEY I. TO LUBEC

By W. O. Smith and J. E. Upson

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IN THE VICINITY OF EASTPORT, MAINE: ESTES HEAD TO TREAT ISLAND AND DUDLEY ISLAND TO LUBEC

U. S. Geological Survey, Prepared in co-operation with the Corps of Engineers, U. S. Army

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 69, PP. 99-120, 1 FIG. JANUARY 1958

SOME BIOSTRATIGRAPHICAL CONCEPTS

BY CURT TEICHERT

ABSTRACT

Time can be measured by observing unidirectional and nonreversible processes, or processes that proceed with a fixed periodicity. Geologic time is measured by observations of rock features that give evidence of previous existence of unidirectional or periodically repeated processes. Periodically recurrent processes do not provide world-wide time standards. The only natural processes that are of a unidirectional and nonreversible nature and that leave universally occurring testimony in the rocks are radioactive decay and evolution of life. The study of past life forms and their distribution in the rocks (paleontology and biostratigraphy) provides a reference system for determining the order of succession of geologic events from the Cambrian to the Recent. Radioactive dating supplies an approximative reference grid of dates in terms of absolute time.

The biostratigraphic zone, which is an assemblage of rocks containing a defined fossil fauna or flora, is the basic and generally the smallest biostratigraphic unit on which world-wide correlations can be established. Application of biostratigraphy to geochronology is known as biochronology. Different plant and animal phyla change in biochronological value during their development, and at different periods during earth history, different groups of animals and plants attained varying degrees of biochronologic importance. The biochronological standard, based for each period on biostratigraphically important groups of fossils, is known as the *orthochronologic* standard. The basic unit of the orthochronologic standard is the time unit corresponding to the deposition of a biostratigraphic zone.

A multiple biostratigraphic terminology is found unnecessary. Zone is accepted as the basic biostratigraphic unit, moment is recommended for the time unit during which a zone was deposited. Instant is suggested for a true "time plane" in stratigraphy. Biochron is suggested for the duration or survival time of taxonomic units, down to the species.

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INTRODUCTION

Stratigraphy is that branch of the geological sciences which is concerned with the study of rock strata for the purpose of establishing their chronological succession and geographic distribution. The historical development of stratigraphy as a science has followed a logical pattern: (1) recognition and interpretation of physical characteristics of sedimentary rocks with emphasis on *lithostratigraphy* from Steno to Werner (seventeenth and eighteenth cen-

tury); (2) recognition of the existence of a meaningful and orderly succession of fossil faunas and floras in sedimentary sequences and development of *biostratigraphy*[1] since William Smith (about 1815); (3) recognition of contemporaneity of dissimilar rocks and fossil assemblages and subsequent development of the *facies* concept from Gressly (1838) to Mojsisovics (1879). Modern stratigraphy rests securely on these three basic achievements of the human mind.

Discussion of stratigraphical concepts in recent years has been much concerned with lithostratigraphy and with facies. These fields have been well covered in modern literature and will not be considered here. Much of the discussion, however, has by-passed or ignored biostratigraphy. If stratified rocks contain fossils, biostratigraphical methods should be applied to their study, in addition to rock-stratigraphical methods. Therefore, the study of fossiliferous stratified rocks is concerned with three aspects: (1) their division into locally mappable rock units, (2) the local sequence of fossil assemblages in the rock units, and (3) the correlation of the rocks through study of their contained fossil assemblages.

The field geologist is often inclined to think of fossils as separate from rocks. However, fossils are not organisms. They are constituent parts of, or structures within, a rock, and their description is an essential part of the definition of the rock unit itself in which they occur. Such a description should include preservation, abundance, morphology (size, shape, etc.), and nomenclature. "Mapping on fossils" is, therefore, a legitimate procedure in geology. In a thick sequence of uniform and monotonous lithology, a single bed may be distinguished by fossil content and hence constitute a mappable unit. Such a bed may be essentially a rock-stratigraphic or a biostratigraphic unit, or it may be both.

World-wide correlation is impossible without the study of the fossils. It is achieved by identifying biostratigraphic units, and biostratigraphic concepts must therefore have status in stratigraphic nomenclature besides rock and time-rock concepts.

Biostratigraphy is concerned with the details of vertical and horizontal distribution of fossils in layered sequences. Hence, biostratigraphic nomenclature should be a system of terms for rock units as characterized by particular fossils. A biostratigraphic unit, therefore, is a bed, or a group of beds, without regard to lithology, containing defined fossils or fossil assemblages. Many geologists have failed to distinguish between, on the one hand, the rocks in which defined fossil assemblages occur and, on the other, the survival times of the assemblages or their individual constituents. Time terms based on survival of taxonomic units can hardly become part of a system of formal stratigraphic nomenclature, but it is necessary to consider some of them here.

The present paper is concerned with biostratigraphy in a most general way. It investigates the basic contribution of biostratigraphy to the system of geochronology, and it ignores the role of biostratigraphy in the solution of local, for example intrabasin, correlation problems.

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Biostratigraphy, Biochronology, and Geologic Time

Time and its Measurement

Time, in the words of Sir Harold Spencer Jones (1950) "... can be thought of in the Newtonian sense as something which flows uniformly". Geologists do not have to concern themselves with the theoretical and philosophical aspects of the time concept. As Lecomte de Noüy pointed out (1937), the Newtonian "uniform flow" is an ideal limit drawn from the actual acceleration or slowing down of the motions of material things. The time with which geologists are concerned is "perceptual time" in the sense of Lecomte de Noüy. For the purpose of the present analysis,

[1] The term (*biostratigraphie*) seems to have been introduced by the Belgian paleontologist Dollo in 1904 (*fide* Diener, 1925), who apparently proposed it in a wide sense for the "entire research field in which paleontology exercises a significant influence upon historical geology." Thus defined biostratigraphy would today embrace paleogeography, paleoecology, paleobiogeography, and other branches of historical geology. Later, Dollo (1910) redefined biostratigraphy as "paleontological stratigraphy." Modern usage of the term as "stratigraphy with paleontological methods" has become more firmly established since Wedekind (1916) and Andrée (1926).

it may be advantageous to distinguish between contemporaneous time and past time. The former may be defined as the time we experience ourselves; the latter is the time that has passed before our personal experience and of which we have only indirect evidence. Yet, the problems involved in the measurement of both kinds of time are fundamentally alike.

In the words of Jones (1950), "Any simple natural phenomenon which obeys one definite law without perturbation might be used to mark off equal intervals of time and therefore serve as a clock". Contemporaneous time is measured by observing and recording phenomena which proceed either unidirectionally and nonreversibly or with a fixed periodicity (movement of a pendulum; rotation of the earth; electrically induced oscillations of quartz crystals; atomic and molecular vibrations). Historical time is that part of the time of the past which has been measured by application of methods similar to some of those now in use. Before that lies geologic time. In order to measure and compare intervals of geologic time, it is necessary to make deductions from observations of natural objects and conditions which may be interpreted as having resulted from the operation of similar factors, that is, from processes which are either unidirectional and nonreversible or have a fixed periodicity. The divisions of geologic time are, therefore, derived concepts, based on observations of rocks. No geologic time divisions could be defined unless it is assumed that processes had been in operation that led to the formation of rocks or minerals from which the nature of these processes could be deduced. A geologist on the moon could not fix and define units of geologic time on the basis of observations of lunar surface geology, because as far as is known, no geologic processes, especially water erosion and sedimentation, similar to those most commonly observed on earth, are active on the moon, and no life has ever been present on it.

Measurement of Geologic Time

Periodic events that may leave recognizable traces in the geological record are the rotation of the earth, the revolution of the earth around the sun, and the various perturbations of the earth's orbit. Among phenomena that result from events of this kind are tree rings and varves (daily, seasonal, or annual) and others of more doubtful interpretative value (*e.g.*, accretion zones of invertebrate skeletons, stratification caused by tides). We may also recall attempts to recognize major climatic cycles (11 years, 33 years, and so on) in sedimentary sequences and attempts to construct curves of climatic fluctuations during the last few hundred thousand years, based on the known perturbations of the earth's orbit (Milankovich's and Zeuner's "Absolute Chronology" of the Pleistocene).

The application of these methods leads to measurements of restricted time intervals in localized areas. They are inherently of geographically restricted validity and will not be discussed here. World-wide effects are claimed for changes in solar radiation as calculated by Milankovich, but even those who support his conclusions see little possibility of projecting the method back beyond the span of 600,000 years, about one five-thousandth of the total length of geologic time, and about one-thousandth of Cambrian to Recent (Phanerozoic) time.

Three major groups of phenomena result from unidirectional and nonreversible processes and leave their testimony in the rocks: (1) accumulation of sediments and related processes; (2) radioactive decay; (3) evolution of life.

Observation of the results of sedimentation processes is limited to restricted areas, and attempts to measure geologic time by the sedimentation method have yielded a limited measure of success only when other methods were used as well (Schuchert, 1931). Among all the estimates of the length of geologic time based on the rate of sedimentation method the one made by Goodchild (1897) stands out as comfortably close to presently accepted figures; Goodchild gives the total time since the beginning of the Cambrian as 704×10^6 years. However, since other contemporaneous estimates, for example by Sollas (1909), were as low as 35×10^6 years, the inconsistencies of these results must have been disturbing.

Estimates of the age of the oceans on the basis of their assumed gradual salinity increase rest on many assumptions, and even if the basis for such estimates could be perfected, they would not help to date geological events other than the time of formation of the first water in the ocean basins.

Many diagenetic processes in sediments are directional and essentially nonreversible, but the rate at which they proceed is known to vary from place to place. For example, fluorine fixation in fossil bones seems to proceed uniformly in any one restricted place and can be used to determine the relative ages of bones

found close together in the same deposit ("Piltdown Man"!), but criteria for absolute dating by this method are uncertain.

The radioactive and paleontologic (or biostratigraphic) methods alone are concerned with objects that are universally available and whose observed physical changes in time may be interpreted as resulting from the operation of unidirectional and irreversible processes which took place on a world-wide scale and proceeded at a more or less uniform rate. Paleontologic analysis, basing its conclusions purely empirically on the little understood workings of living matter, supplies the general sequential pattern of geologic events. Study of radioactive processes furnishes an approximation to an absolute time scale in terms of years; its disadvantages are a comparatively wide margin of error in most measurements and scarcity of suitable material on which measurements can be made. The literature contains few age determinations (perhaps no more than one) on syngenetic radionuclides from paleontologically defined stratigraphic units (*see* Moore, 1951), and almost all radioactive age determinations are made on igneous, hydrothermally introduced, or secondarily transported minerals that cannot as a rule be referred to a precisely defined place in the stratigraphic succession. At present no coherent picture of the history of the earth could be built on the basis of radioactive datings.

The major shortcoming of the biostratigraphic method is its inability to furnish an absolute time scale. Its main advantages over all other methods are that its material is fairly abundantly available and that it is based on established criteria by which the succession in time of physical events may be judged and their time relationships in distant parts of the world determined.

I do not wish to suggest that in the biological world genes mutate with the predictable precision of radioactive decay. Evolution is continuous only in the sense that life itself is a continuously operating chain of events. There are discontinuities and jumps, and it is these that are often particularly useful and valuable in the paleontologic record. However, the over-all effect of evolutionary processes which determine the composition and character of life communities was felt all over the world at the same time and at the same rate. On this uniformity of the over-all result depends paleontologic correlation, a highly empirical science that can never take its own results for granted and yet has succeeded in accumulating a huge body of almost infallible guides to the interpretation of the geological past. These are known as "index fossils" or "guide fossils". Spieker (1946, p. 146) emphasizes the empirical nature of index fossils in the following words:

> "First, there is nothing inherent in fossil occurrences that will serve without question as indication of contemporaneity. Second, when all possible conditions of occurrence and interpretation are evaluated and a balance of probabilities struck, fossils afford the surest means yet available for correlation."

Arnold (1948, p. 370) has rightly stated that "a guide fossil remains a guide fossil only until it has been discovered at some place where it is not supposed to exist".

Moore (1952) has also stressed the empirical nature of the biostratigraphic evidence. He refers to uneven distribution of populations on the sea floor, the influence of sedimentary environment, distinctions between faunal provinces, and large disparity in conditions of fossilization as some of the most important factors that "introduce complexity of spatial distribution of the fossils".

On the whole, the geologist seems to be no worse off than the physicist, for, in the words of Oppenheimer (1953, p. 24),

> "No experimenter takes his equipment quite so much for granted that he fails to check whether in fact it is performing as it is supposed to perform; but the notion how it is supposed to perform is for him in general a fixed thing, not calling for further inquiry."

As an example, Oppenheimer mentions the photographic plate which ". . . has served as an instrument of science for decades, during which its behavior was only very incompletely understood". The analogy to our problems is obvious. In paleontologic correlation the geologist has a general notion of how fossils are expected to perform, but he can never relax his vigilance in checking whether in fact they are performing as they are supposed to perform. Insufficient knowledge of paleontologic facts and incorrect interpretation of the paleontologic record have led to, and are still responsible for, grave errors in the interpretation of historical succession (*i.e.*, correlation), but this fact does not discredit the method.

Significance of Biochronology

The application of paleontological studies to stratigraphic problems is known as *biostratig-*

raphy, and the dating of geologic events by biostratigraphic methods is known as *biochronology*. Biostratigraphic investigations are basic to all time and time-rock concepts and are the only methods of establishing time relationships of rocks deposited in different parts of the world, whereas in closely adjacent areas other methods may also be applied. The only sure way of recognizing and comparing, on a world-wide scale, time-rock units of any order of magnitude, whether systems, series, or stages, is on a paleontologic basis.

Biochronology shows the order of succession in which events occurred; a loosely fitting absolute scale comes from the radioactive methods. But in geological studies it is the order of events that is of first importance, not their duration in terms of years. Also, between widely spaced radioactive dates biochronology can, by interpolation, supply more accurate smaller absolute time intervals than any known radioactive "chronometer". Radioactive dates still have to be carefully checked against biochronologic information, and if the discrepancy seems too great, the tendency is to discredit the radioactive rather than the biochronologic data.

Stratigraphy and historical geology were well established as sciences and had developed a coherent and logical picture of earth history long before absolute time measurements were possible. This situation is best illustrated by the fact that all geological systems which are now internationally recognized, with the exception of the Ordovician (which arose as a compromise solution of an old boundary problem), were defined, named, and their time sequence established, within 3 decades of the promulgation of William Smith's "law of the orderly succession of fossil faunas in stratified rocks", and about 70 years before the first series of radioactive dates became available.

Spieker (1956) seems to deplore the "freezing" of the geological time scale at about this time. On the contrary, I would regard this situation as a natural consequence of William Smith's discovery. It is the measure of a natural law that its discovery relates the previously unrelated. Radioactive dating during the past 50 years has not required any change in the general stratigraphic scheme as established on paleontologic evidence by the mid-nineteenth century. It is not at all important that the names and boundaries of the geologic periods and the lengths of the eras would have been different, if stratigraphy after William Smith had developed not in Europe, but in America, India, or eastern Asia. The general space-time system of reference would necessarily have been developed in the same way and would have provided the same scale. Similarly, if the O-Meridian were shifted to some place other than Greenwich, the geographical co-ordinates of every place on earth would be changed, but not its position in space.

The picture of earth history that had been assembled solely on the basis of paleontologic correlation may be likened—on another plane—to a map without a scale. If we had to use such a map, we would somehow have to work out a rough scale. Naturally the map becomes more meaningful as soon as a more nearly accurate scale is supplied.

Biochronology had succeeded in determining the relative lengths of geological periods with a considerable degree of accuracy before the advent of radioactive measurements. Perhaps the last exhaustive and serious review of the problem of the absolute length of geologic time from the paleontologic viewpoint, before the discovery of radioactivity, is in a long-forgotten book by H. S. Williams (1895). After discussing and comparing the various physical methods then in vogue for the measurement of geologic time—resulting in estimates anywhere between the Biblical 6000 and McGee's 14 billion years for the age of the earth—Williams proceeds to an estimate of the relative duration of the geological periods solely on the basis of interpretation of the fossil record. His results are astonishing. Without committing himself to absolute figures, Williams suggests that the Paleozoic is equivalent to 65 per cent, the Mesozoic to 20 per cent, and the Cenozoic to 15 per cent of the time that has passed since the beginning of the Cambrian. Present best estimates based on radioactive methods are 64.5 per cent for the Paleozoic, 24 per cent for the Mesozoic, and 11.5 per cent for the Cenozoic.

The Orthochronological Standard and its Place in Geochronology

Basis of Orthochronology

Biochronology is the only branch of geochronology that supplies a sufficiently dense net of control points in a space-time system of co-ordinates to allow the construction of a reasonably logical and integrated picture of the history of the earth since the beginning of the

Cambrian. The rich detail of knowledge of the Paleozoic, Mesozoic, and Cenozoic (that is, the Phanerozoic) eras in contrast to the relatively confused historical picture of the Precambrian (or Kryptozoic) is proof of the fundamental role of biochronology in geochronologic interpretations. The purpose of this section is to investigate briefly the nature of the biostratigraphic tools with which the edifice of biochronology has been erected and is constantly being improved.

In contemporary writing, there is a tendency to neglect or to discredit the value of the biologic species and genus in stratigraphic correlation. This is perhaps partly due to the decreasing emphasis on paleontologic training among field stratigraphers and, therefore, increasing lack of appreciation of basic paleontologic methods among this group. Increasing complexity of paleontologic nomenclature also contributes to lack of understanding between paleontologist and geologist. No easy solution of this difficulty is in sight, and proposals to restrict arbitrarily the number of generic and specific names in paleontology (Weller, 1952; Schindewolf, 1954) seem to be lacking in realism, if it is remembered what a small proportion of the potentially available fossil species has been described (Teichert, 1956). A recently published symposium (Sylvester-Bradley, 1956) represents an important step toward recognition of the many problems peculiar to the study of "paleospecies".

In spite of inherent weaknesses the paleontologic species and genus continue to be the most reliable, least ambiguous yardsticks in interregional stratigraphic correlation, if used with circumspection and all possible safeguards. The survival times of species and genera vary greatly, from the time span of a subzone (perhaps less than half a million years in some mammal groups, and almost certainly even less in certain graptolite and ammonoid lineages) to the time span of a system or more (as much as 30 to 300 million years) in some brachiopod genera such as *Atrypa* or *Lingula*. The stratigraphic paleontologist sifts and weighs the fossil material and selects, for chronological purposes, those species which are most suitable and reliable in any one period, epoch, or smaller time division over the greatest possible geographical area. This task is entirely empirical. Any group of animals or plants may somewhere, at some time, furnish infallible zone markers. For example, although crinoids generally do not rate highly in paleontologic correlation, the genera *Uintacrinus* and *Marsupites* are important zone fossils of universal distribution during a short interval in the Upper Cretaceous, and the Permian sequences in geographically separated sedimentary basins in Western Australia can be zoned by a succession of species of the crinoid *Calceolispongia* (Teichert, 1949).

The accumulated experience of generations of paleontologists has shown, however, that every geologic period has its characteristic groups of short-lived species or genera and that the stratigraphic value of biologic groups varies greatly from system to system. Although Foraminifera have existed at least since Ordovician time, they are useful in world-wide zoning and correlation only in sequences of Pennsylvanian, Permian, Late Cretaceous, and Tertiary (up to Miocene) age. Trilobites lived throughout the Paleozoic, but only in the Cambrian does the distribution of many of their species afford a base for wide regional, even world-wide correlation (Kobayashi, 1949; Lochman-Balk and Wilson, 1957, Am. Assoc. Petroleum Geologists Program, 42d Ann. Meeting, p. 48). On a more restricted continental basis trilobite species are still used as zone fossils in the Ordovician, and their value in later periods diminishes rapidly. But even in the Middle Paleozoic the biostratigraphic value of at least some trilobites is unimpaired, as illustrated by the world-wide survival of *Harpes* and *Scutellum* into the Frasnian (basal Upper Devonian).

Stratigraphic usefulness of a major biologic group is not necessarily a function of its universality and abundance in sedimentary rocks. Gastropods and pelecypods have been common in faunal communities all over the world since the early Paleozoic but are rarely used as zone fossils before the Tertiary. A comparatively short-lived and little diversified group like the Graptolitoidea yields dozens of the most important index species in the Ordovician and Silurian, whereas their closest relatives, the dendroids, with one or two exceptions are of no use to the stratigraphic paleontologist. Figure 1 represents an attempt to illustrate the degree of stratigraphic usefulness of important groups of invertebrates in the geologic column. (*See* also Simon, 1948.)

Marine invertebrates are generally most useful for the setting up of correlation standards of world-wide validity, because the environment in which they live is less exposed to sudden changes and is of greater uniformity than the continental environments. It should be remembered that, contrary to widely held

beliefs (Dufour, 1951), whether a marine species is benthonic or planktonic is not of great importance as far as biochronologic usefulness is concerned. In a suitable environment benthonic animals may disperse with sufficient rapidity to make them useful zone fossils. The value of planktonic forms is that they may be transported across, and may die over, many different kinds of bottom environments. Hence, as fossils, they transgress facies boundaries and may help to tie in two or more biostratigraphic units based on biota controlled by different bottom environments.

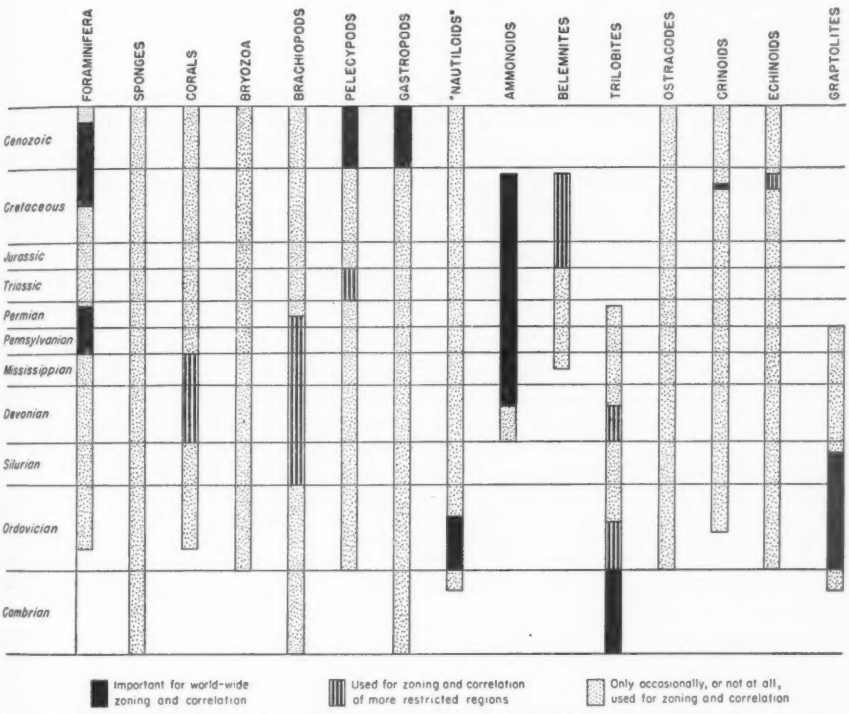

FIGURE 1.—RELATIVE BIOCHRONOLOGIC SIGNIFICANCE OF THE MAJOR GROUPS OF MARINE INVERTEBRATA DURING GEOLOGIC TIME

Figure 1 shows at what periods some of the major phyla were stratigraphically important and what groups are used for the orthochronologic standard during any one geologic period. Thus at the end of the Cambrian period the prevailing trilobite chronology is replaced by a graptolite chronology. Ordovician graptolite-bearing rocks are some of the most easily and most securely datable and correlatable rocks in the geologic record. Where no graptolites are present, considerable difficulties in correlating Ordovican rocks are commonly experienced, except in the Early Ordovician where nautiloids may be helpful. Lower and Middle Devonian rocks are not so accurately correlatable as Upper Devonian strata where ammonoids become abundant, and Müller (1956) has recently claimed that conodonts, in this epoch, may serve as even better index fossils.

These few examples may show how Figure 1 should be read. The representation is generalized, because, given the right time and the right environment, stratigraphically and chronologically useful forms may emerge from almost any group of organisms, though usually of more restricted regional application; examples are fresh-water pelecypods in Carboniferous coal measures, echinoids in the Upper Cretaceous of Europe, fishes and ostracodes in lacustrine and brackish deposits, reptiles in the Triassic of the southern continents, mammals in the Tertiary of the northern hemisphere.

On the other hand, in the same bed, or se-

quence of beds, fossils of very unequal biochronological value may be found. Such associations are well known in local sequences. Occasionally, their recognition and proper evaluation is critical for the dating of world-wide events. Thus, in Brazil Late Paleozoic glacial deposits are associated with, or overlain by, a variety of beds which contain pelecypods and plants of Pennsylvanian aspect. From this Maack (1952), Beurlen (1953), and others have concluded that the Late Paleozoic glaciation not only in Brazil, but in other southern continents, including Australia, occurred during the Pennsylvanian period. In Australia, however, these same plants and pelecypods are either contemporaneous with or younger than typical Permian ammonoids which occur in rocks either associated with or stratigraphically above glacial sediments (Clarke *et al.*, 1951; Teichert, 1954). There is no real discrepancy or conflict here. It is well known that at the Pennsylvanian-Permian boundary neither pelecypods nor plants are of great biochronological value and that locally Pennsylvanian elements of these groups may survive into Permian time. On the contrary, the appearance of typical Permian ammonoids in rocks of Pennsylvanian age would be nothing short of sensational. Application of biostratigraphical experience, therefore, favors an Early Permian rather than a Pennsylvanian age of the Late Paleozoic glaciation of the southern continents.

The succession of species or genera that have been chosen from among the useful groups in each geological period serve to characterize successive biostratigraphic zones which provide the standard for the relative chronology of that period. Following a proposal by Schindewolf (1950a) the chronology based on a standard succession of biostratigraphically significant faunas or floras should be known as *orthochronology*, because it is based on irreversible evolutionary processes, even in those cases (and they are probably in the majority) where successive zone fossils do not represent members of an evolutionary lineage. The most ideal case of an orthochronological system of reference is a stratigraphical succession of species where each successive species is the descendant of the one which immediately precedes it stratigraphically. Such evolutionary lineages are abundantly known from all animal phyla of all ages, both in marine and terrestrial environments. The Indonesian paleontologist Tan Sin Hok (1939), on the basis of detailed "phylomorphogenetic" studies of Oligocene and Miocene miogypsinids and lepidocyclines, called such series "paleontological chronometers". But this term, though imaginative, is not quite applicable, because a chronometer is supposed to measure absolute time, or at least give a close approximation of it.

Where the guide fossils of successive zones are not linked by evolution, those in the higher beds are descendants of forms which were stratigraphically unimportant in earlier beds. In some cases their origins are unknown. They seem to appear "out of the blue", their ancestors having developed in areas where either no sedimentation took place or where sediments were later removed by erosion or buried too deeply below younger sediments or where they have not yet been found. However, even in zonal successions where successive guide fossils do not form evolutionary lineages their appearance is due to the continuous operation of evolutionary mechanics, which assures that the same life forms are never produced twice. Some problems raised by appearance of successive zone fossils, not linked by evolutionary descendency, have recently been reviewed by Westoll (1954) on the basis of paleogeographic and facies considerations. In this connection it should also be remembered that the irreversibility of evolution is a natural law only in so far as it is simply an expression of the very high mathematical improbability that a biological species or genus should appear more than once in the history of life on earth. The number of gene mutations and their combinations is so great and the influences of natural selection so manifold that the repetition of identical gene combinations in identical environments can be discarded as a practical possibility. (*See* Teichert, 1943, p. 175; Schindewolf, 1950c, p. 209.)

Validity of Orthochronologic Standard

Introduced into or originating in a given suitable environment, a species will spread with great rapidity as far as the limits of that environment. Beyond those limits it will never advance. The Recent Indo-West-Pacific marine faunal province is a shallow-water community of extraordinary uniformity that extends along tremendous stretches of coast line from Durban on the east coast of Africa, to India, Indonesia, northern Australia, southern Japan, and to the Hawaiian Islands, a total distance of about 12,000 miles, and occupies several hundred thousand square miles of littoral and neritic environment. Zoogeographers recognize subregions in this province, but there are many

species common to all of them. This faunal province is recognizable in deposits of Pleistocene age over the same area, yet in spite of its age and its enormously wide distribution in the tropical seas, the fauna has been unable to surmount certain barriers. It has never rounded the Cape of Good Hope, because of the prevailing cold currents along that part of the coast of Africa, and the fauna of the tropical Atlantic is quite dissimilar. Also, the landless voids of the eastern Pacific have proved an almost impenetrable ecologic barrier, and the Pacific faunas of tropical Central America are more closely related to those of the West Indies with whom they share common ancestors in the Tertiary seas. Marine sediments that are being deposited today in this zoogeographic province form a biostratigraphic zone *in statu nascendi*, with an abundance of remains of characteristic species and genera. In future geologic periods this zone will be easily traceable halfway around the world and will provide the material basis for one of the best documented chronological units of the Quaternary period. It would seem that uses and limitations of the index-fossil method can be fully illustrated by study of present-day animal and plant geography, most of all by study of the zoogeography of the present-day oceans (Ekman, 1953).

The question then arises: How good do our biochronological yardsticks have to be in order to enable us to develop a reasonably accurate picture of the succession of geologic events? The ultimate aim is reconstruction of successive paleogeographies on paleogeographic maps. What length of time does a paleogeographic map have to represent to give a reasonably meaningful picture of the distribution of land and sea and of faunas and floras at any one time? Maps depicting, for example, the "paleogeography of the Ordovician" are somewhat misleading. They represent the distribution of Ordovician rocks rather than give a picture of the paleogeography at any one time during the Ordovician period. The amount of detail that can be shown on a paleogeographic or paleobiogeographic map depends on the length of the time interval which it attempts to represent.

The Quaternary period, about 1 million years long as determined by various means, may serve as an example. One million years ago the margins of the continents were essentially the same as now, the mountains must have had about the same heights, the drainage systems did not differ very much from those of today, and the same rivers discharged into the same seas. No violent orogenies occurred during this interval, although there have been secular vertical movements, mainly caused by glacio-isostasy. The most disturbing events during the last million years were caused by extraordinary climatic fluctuations which cannot be regarded as typical of most former geological periods. In the absence of these climatic fluctuations there would have been, during the past million years, no major eustatic sea-level changes, and isostatic movements of the continents would have been on a lesser scale. Over-all composition of marine invertebrate faunas has changed little during this time, although there were shifts in the boundaries of zoogeographic provinces (biofacies). Changes in the composition of land faunas, particularly vertebrates, have been more noticeable, and a greater percentage of species failed to survive.

If earth history of the last million years is any guide, a paleogeographic map of the world probably cannot represent conditions for a shorter span than approximately 1 million years. The same goes for a general characterization of the flora and fauna of the earth. Actually, during long periods of the earth's history paleogeographic, faunistic, and floristic changes may have been so slow that paleogeographic and paleobiogeographic maps showing conditions several million years apart would reveal no significant differences.

According to present knowledge it would seem that biostratigraphic zones in different geologic periods are not entirely equivalent in length of corresponding absolute time. The time equivalent of a trilobite zone in the Cambrian can be estimated to be about 2½ million years, that of a graptolite zone in the Ordovician possibly as much as 5 million years, that of some ammonite zones in the Jurassic no more than 300,000 years. (*See* Kobayashi, 1944; Schindewolf, 1950b; Jeletzky, 1956.) An interval equalling a "zone time" is probably the shortest interval for which we can hope to give a reasonably accurate representation of conditions on the earth, except for certain episodes in the most recent history of the earth, and this time interval is of the general order of between half a million and 5 million years. This is of the general order of Gilluly's (1949) estimate of 3–4 million years for the minimum recognizable and correlatable time unit in geologic history.

That such a time span is more than ample to allow most biologic species to occupy the earth to the limits of existing suitable environments has been abundantly demonstrated by examples of the dispersal of species accidentally

transported into new fitting environments. A classical case is that of the migration of the marine gastropod *Littorina litorea*, which was introduced from Europe to Nova Scotia about 1870 and, within 40 years, spread along 750 miles of coast line as far as New Jersey. Schindewolf (1950b) pointed out that at this rate a species could have migrated at least 500 times around the world during a time span equivalent to the deposition of a single biostratigraphic zone. The rabbit and the *Opuntia*, once introduced in Australia, populated large areas of the continent with extreme rapidity. The muskrat (*Ondata zibethicus*), after having been introduced to Europe from North America in 1905, colonized an area of 64,000 square miles within 28 years, and it now occupies huge areas in western, central, northern, and eastern Europe and in Siberia. De Voos, Manville, and van Gelden (1956) have recently described many examples of the spreading of mammals in recent times. Kurtén (1957) has given further examples from the fossil record and estimates that the unchecked spread of a mammal species may proceed at a pace of about 1000 km per century. At this rate a mammal species could spread around the world in about 4000 years. Relevant data have also been discussed by Simpson (1940), Deevey (1949), and H. and G. Termier (1956).

To represent the changing geographies and biogeographies of the world during geologic time since the beginning of the Cambrian a sufficiently accurate picture could be obtained from a series of 150 to 200 paleogeographic and paleobiogeographic maps. This is only five or six times the number of the most extensive series of combined paleogeographic and paleobiogeographic maps of the world published to date (Termier and Termier, 1952).

Only biochronology provides the basis and general framework for this sequence of pictures on which generations of stratigraphers and paleontologists have worked in the past and will continue to work in the future. This is biochronology's major contribution to geochronology. Other geochronologic methods may provide much fine local detail and greater physical accuracy, most of which, however, would be of little scientific value, unless it could be referred to a place in a biochronologically determined system of reference.

The Zone Concept

Definition

The fundamental unit in rock stratigraphy is the *formation*. Through definition and description of successive formations, a picture of the physical history of an area is built up. The fundamental unit in biostratigraphy is the *zone*. A biostratigraphic zone is a bed or group of beds which contains a defined assemblage of fossils, one of which is selected for the naming of the zone. The wider the horizontal distribution of these assemblages or some of their constituent parts, the greater the value of the zone as a biochronological marker. The succession of zones gives the biochronological order of the events in the world of living organisms. Both formation and zone are empirical concepts whose definition must be adapted to the needs of the situation.

The custom of naming a biostratigraphic zone after one of its contained fossils has at various times given (and is still giving) rise to the following misconceptions: (1) the name fossil is restricted to the zone that is named after it; (2) only the name fossil, and no others, occur in the zone; (3) all rocks in which the name fossil occurs belong to the zone; (4) rocks in which the name fossil does not occur may not be included in the zone.

In fact, a biostratigraphic zone is defined by an assemblage of species or genera, rather than by one species or genus; the name fossil may also be a member of earlier or later fossil assemblages; and it may in places be absent from the typical assemblage which characterizes the zone.

In 1950, I pointed out the fact that many stratigraphers use the term zone for time units. With Diener, Arkell, Jeletzky, and many others, I maintained that the term had been introduced by d'Orbigny and more firmly established by Oppel for rock units characterized by distinct fossils or fossil assemblages, and I suggested that this definition places the zone concept close to the time-rock category, based on biostratigraphical criteria. This statement needs modification, and since there is still no unanimity of opinion in this matter, it is advisable to present this historical background more fully.

Historical Review

Notwithstanding Schindewolf's (1950b; 1954; 1957) and Fiege's (1927; 1951) assertions to the contrary I can find nothing in either d'Orbigny's or Oppel's writings to suggest that these authors used, or intended to use, zone as a time term.

In 1850 (p. 602) d'Orbigny divided the Jurassic rocks of France into 10 "*zones ou étages*". That he sometimes used the two

words synonymously is clear from his later *Cours élementaire* (1852). However, in volume 2 (p. 282) of that work, he also writes of the "ordre chronologique des étages, en zones superposées, distinctes, constituant des étages", which suggests that d'Orbigny recognized zones as subdivisions of *étages*. This interpretation is supported by another passage (1852, p. 440) of the same volume where he proposes to include in the "étage sinémurien toutes les couches jurassiques inférieures à la Zone de *l'Ostrea arcuata* qui sert toujours et partout de dernière limite superieure à cet étage". In other connections, however, d'Orbigny uses "zone" and "étage" synonymously. He was not entirely logical in his separation of time and rock concepts, and in a few places he used *étage* in a time sense; thus in 1852 (p. 256): "un étage, pour nous, est un époque en tout identique à l'epoque actuelle. C'est un état de repos de la nature passée . . ." This sentence, of course, expresses d'Orbigny's general geological philosophy of successive creations followed by long periods of stability and rest, but wherever he speaks of *étages* in the descriptive parts of his works, he means rock units characterized by a distinct fossil fauna.

The zone concept was more definitely established by Oppel (1856–1858). As Arkell (1933) remarked, Oppel never gave a formal definition of the term, but the careful reader of his "*Juraformation*" will be in no doubt that he used the term for rock units, not in a time sense. Oppel divides the Jurassic into *Etagen* which coincide to some extent with d'Orbigny's *étages*. The decisive step beyond d'Orbigny is Oppel's consistent division of the *Etagen* into zones. He, therefore, calls his Etagen *Zonengruppen* (groups of zones). A zone, for Oppel, is any bed in which he can recognize a defined fossil assemblage, irrespective of the lithology of the bed. It seems impossible to interpret Oppel's intentions in any other way, for throughout the book he writes of thicknesses of zones, of overlying and underlying, higher and deeper, zones; he describes the lithology of the zones in various areas, and "outcrops of zones" are frequently mentioned. A few quotations may suffice:[2]

"Ich nannte . . . diese oberen Schichten vorläufig Zone des *Amm. geometricus* . . ." ("I called . . . these uppermost beds tentatively the zone of *Ammonites geometricus* . . .") (1856, p. 157).

". . . die Zone des *Amm. Jamesoni* mehr als 100 Fuss mächtig . . ." (". . . the zone of *Ammonites Jamesoni* more than 100 feet thick . . .") (1856, p. 241).

"Obschon die Zone der *Posidonomya Bronni* nicht überall durch Schiefer gebildet wird . . ." ("Although the zone of *Posidonomya Bronni* does not everywhere consist of shale . . .") (1856, p. 317).

"Im Depart. Calvados liegt die Zone des *Amm. Macrocephalus* über dem dortigen Cornbrash, steht aber in mineralogischer Beziehung in enger Verbindung damit." ("In the Department of Calvados the zone of *Ammonites Macrocephalus* lies above the Cornbrash in that locality, though the two are mineralogically closely similar".) (1857, p. 216).

". . . die Zone des *Amm. anceps* ist besonders zu Oberlemmingen . . . deutlich aufgeschlossen." (". . . the zone of *Ammonites anceps* is particularly well exposed near Oberlemmingen.") (1857, p. 223).

"Ich nenne die Zone, welche durch die mittleren Oxford-schichten gebildet wird: Zone des *Amm. biarmatus* . . ." ("I am calling the zone which consists of the middle Oxford beds, the zone of *Ammonites biarmatus* . . .") (1857, p. 314).

"Die Bonebedschichten habe ich . . . Zone der *Avicula contorta* genannt." ("The Bonebed strata I have called . . . zone of *Avicula contorta*.") (1858, p. 248).

In Oppel's final table (1858, p. 256) there is one column headed "Zonen (Lager oder Stufen, d.h. paläontologisch bestimmbare Schichtenkomplexe)." A more precise definition of zone would be difficult to find. This passage leaves no doubt that Oppel's zones were "paleontologically identifiable complexes of strata".[3]

Among those who have come to the same interpretation of Oppel's works are Diener (1919), Arkell (1933; 1956), and Kleinpell (1938). "Zone" (ζωνή) in Greek meant "belt", and it is hard to understand how a word that in any language has a spatial connotation should have come to be used for divisions of time. Nobody divides history into zones, and the concept of "time zones" (except as used for longitudinal belts of standard times on earth) seems illogical.

[2] These quotations were selected some years ago. In the meantime Arkell (1956) has published his own selection of passages from Oppel's works in support of the same thesis. The fact that there is no duplication in our lists may be taken as confirmation of Arkell's opinion that statements of Oppel's views, as Arkell and I interprete them, may be found in "countless other places throughout his books."

[3] Kobayashi's (1944) suggestion to introduce the term "orthozone," for a zone in Oppel's sense as here understood, seems to have little to recommend it. This term would be subject to the same misinterpretations as the original "zone."

Zone and Time

Almost universally those who have recognized the zone as a spatial rather than a time concept, including myself (1950), have regarded it as part of the time-stratigraphical system of nomenclature. (*See* Schenck and Muller, 1941.) However, the zone is a true biostratigraphic concept. The difference is this: A time-stratigraphic unit comprises all rocks deposited during a given time interval; correlation may be made by many methods, so that an element of subjectivity cannot be eliminated from the definition of time-stratigraphic units. Biostratigraphic units contain identical or closely comparable faunas and floras and are, therefore, objectively correlatable. Reliability of correlation of time-stratigraphic units increases with the number of biostratigraphic units which can be incorporated in them. This was clearly recognized by the American Commission on Stratigraphic Nomenclature (1952), which, however, might well have stated with greater clarity and emphasis the role of biostratigraphy in accurately defining the boundaries of time-stratigraphic units.

Tomlinson (1940) and Hedberg (1941; 1954) justly advocated separation of biostratigraphic and time-stratigraphic nomenclature. However, both failed to recognize the true time significance of biostratigraphic units in the sense here discussed.

The Second International Geological Congress at Bologna, 1881, as well as some of the subsequent congresses, devoted much time and effort to clarification and unification of stratigraphic nomenclature. What uniformity in international usage now exists is largely the result of efforts of the committees of these first congresses. The zone concept, however, received scant attention, and in the final recommendations of the Bologna congress (Dewalque, 1882) the zone was somewhat offhandedly defined as "un ensemble de couches d'un ordre inférieur, caracterisé par un ou quelques fossiles spéciaux, qui servent a la dénommer." Fifty-one years later an almost literal translation of this passage was made part of the "Ashley Code" (Ashley *et al.*, 1933, p. 429). This meager statement, however, does not seem to do justice to such a basic concept. Also, as Tomlinson (1940) pointed out, there are inconsistencies in the definition of zone in different parts of the "Ashley Code".

Especially in German paleontologic literature, the concept of zone gradually changed into one of time connotation. (*See* Diener, 1925; Arkell, 1933; Schindewolf, 1950a.) As now widely used in Germany and elsewhere the term zone signifies the survival time of a species or other taxonomic group. Schindewolf (1954, p. 32) maintains that the zone is a pure time concept and illustrates this by saying that, for example, the zone of *Macrocephalites macrocephalus* is represented also in places where this species or its associated fauna does not exist. Such usage may easily lead to conflicts and confusion, where a multiple zone terminology is available. Thus in the lower Permian there is a choice between the fusuline zone of *Pseudoschwagerina* and the contemporaneous ammonoid zone of *Properrinites*. Both are useful biostratigraphic divisions. But there is no advantage in a terminology which tells us that the zone of *Pseudoschwagerina* also exists where only ammonoids occur, or that the zone of *Properrintes* should embrace a pure fusulinid facies. Such statements would not seem meaningful, because the two zones replace each other in different environments and complement each other in the biochronological scale.

In the Cretaceous, some ammonoid zones as well as zones based on planktonic Foraminifera (*Globotrucana*) have world-wide validity and applicability. If zones were to be regarded as time units, and since there can be only one name for any one time unit, which of these standards should we use?

Examples may be found in every system. The Devonian, according to the prevailing biofacies, may be biostratigraphically zoned by corals, brachiopods, or ammonoids (goniatites). With the help of the combined and integrated evidence of these three zonal systems, almost any sequence of Devonian age anywhere in the world can be correlated and dated with a high degree of accuracy. Only some of the coral and brachiopod zones, but almost all the goniatite zones, are of world-wide validity.

As an example of the latter consider the goniatite genera *Manticoceras* and *Cheiloceras*. *Manticoceras* appears with a considerable range of species and in great abundance of individuals in the beginning of the Upper Devonian and is of the world-wide distribution in rocks of Frasnian age. Rocks with *Manticoceras*, whether in North America, Europe, central Asia, Morocco, or Australia are assigned to the Frasnian Stage. Where sedimentation is continuous into the higher Upper Devonian, *Manticoceras* disappears at a certain level, and, if ecologic conditions continue to remain favorable to the existence of goniatites, species

of an altogether different genus, *Cheiloceras*, appear. These indicate the base of the Famennian Stage. Both *Manticoceras* and *Cheiloceras* are, in their respective strata, accompanied by other genera, some of which (for example, *Tornoceras*) are common to both faunas. Thus, *Tornoceras* is less suitable as an index fossil than either *Manticoceras* and *Cheiloceras*, which have never been found together. A century and a half of research in Europe has shown them to occur in separate layers, *Cheiloceras* above *Manticoceras*, never in reversed succession, and later investigations in other continents have led to the same results. (*See* Miller, 1938; Teichert, 1943.) The succession of these genera can, therefore, be used to determine sequence in layered rocks. They are two small but solid bricks in the structure of orthochronology.

As to the time relationships, it would be nonsensical to assert that the zonal boundary between rocks containing *Manticoceras* and rocks containing *Cheiloceras* "transgresses time". Such a hypothesis would require assumption of a highly unlikely pattern of faunal migrations, where swarms of species of *Manticoceras* are followed, everywhere at the same distance and the same time interval, by swarms of species of *Cheiloceras*, the two waves preserving their separate identities on a staggered mass migration around the world, possibly throughout millions of years, without evolutionary changes and without ever becoming mixed. This picture is unreal. The only realistic conclusion is to assume that the boundary between the *Manticoceras* and the *Cheiloceras* zones is a true time plane.

It would be easy to repeat this investigation for almost every critical zone fossil or fauna throughout the geologic column for hundreds, perhaps thousands of test cases. The conclusions would be the same. In the words of Jeletzky (1956) we would have to "invoke a miracle", if, for example, we were to assume anything but world-wide contemporaneous deposition for each of the 55 ammonite zones of the Jurassic. Not all of them occur everywhere, but wherever two or more are found in superposition they occur in the same order. Arkell (1957, p. L112) summarized the picture of the stratigraphic and regional distribution of the Mesozoic ammonoids as follows:

"Evolution is above all very uneven. Certain periods were outstandingly productive of new and virile forms which often seem to have sprung into existence from nowhere . . . and to have become dominant almost simultaneously over a large part of the world. These are the periods of *paedomorphosis, macroevolution, saltative evolution, explosive radiation* or evolutionary deployment, according to the terminology of various biologists and geneticists. How such sudden multiple creations were brought about is a task for the future to determine."

Hedberg (1948; 1954), on the other hand, has stated:

"The first specimen of any newly developed species is bound to be older than any other specimen and the first occurrence in one area is never exactly of the same age as the first occurrence in another area. New life forms did not spread instantaneously. . . ."

This statement is correct, if "exactly of the same age" and "instantaneously" are taken to mean what they do in daily life. In geologic writing, however, the standards for contemporaneity and instantaneousness are more flexible. Perhaps H. and G. Termier's (1956) expression quasi-instantaneous (*quasi-instantané*) could be recommended as a convenient term to indicate instantaneousness in the geologic sense.

In a few cases, species (or genera) that were believed to be reliable zone markers occur in a bed "where they are not supposed to exist". This is no proof that such paleontologic zones are time-transgressive. It only suggests that we must either find a better index fossil or redefine the zone.

The stratigraphic ranges of many individual species, genera, or higher taxa are not the same everywhere in the world. However, some of the classic examples for alleged survival of life forms in some areas, long after they have become extinct elsewhere, are based on superficial evaluation of the evidence. *Nautilus* and *Trigonia* are perhaps the most popular examples, although they serve much better as examples for the longevity of textbook errors. The genus *Nautilus s. str.* is unknown from pre-Recent rocks (Miller, 1951), and *Trigonia* became extinct at the end of the Cretaceous. Its living descendant in Australian waters is *Neotrigonia*, and none of the extant species of this genus are known from rocks older than Pleistocene in age. *Latimeria*, the living coelacanth, and *Neopilina*, the even more astounding recently discovered survivor of an early Paleozoic gastropod family (Lemche 1957), are not known as fossils. They are descendants of different forms of orders of fishes and gastropods which were more abun-

dant in earlier periods. There is thus no danger that finds of nautilids, trigoniids, Monoplacophora, or coelancanths in sediments that are being laid down today will seriously upset the biochronology of future generations of geologists.

Once a zonal succession is established in one geologic region, it is not easily upset by stratigraphic observations elsewhere. Zones may be found missing, but the succession will not be found to be reversed or arranged in disorderly fashion. Spath (1931) called attention to certain seeming discrepancies in the vertical succession of ammonite species in the Lower Jurassic sequence of Champfromier, near Bellegard in the Jura Mountains of France, as compared to the classical English zonal sequence. From these reports Spath attacked the "unsoundness of polyhemeral [= polyzonal] Liassic chronology" and argued that

> "disregarding the warning of Forbes, Huxley, and later defenders of homotaxis, we got into the habit not only of measuring all successions by European standards, but of relying on a scheme of zones based on what the distribution of ammonites might have been if arranged specially to our convenience."

It has taken 25 years to solve this biochronological enigma, but after reviewing the evidence Arkell (1956, p. 104) concludes that the seeming discrepancies of the Champfromier section arose from a combination of misidentifications of species and disturbances of the section by landslides.

According to Hedberg (1948, p. 457) "the time-value of faunizones may fluctuate from place to place with faunal facies variation". That facies also determines composition of faunas (and floras) is now commonly recognized. But facies can never lead to a reversal of evolutionary processes. No facies conditions have been discovered that have made it possible for *Didymograptus bifidus* to survive into the time when *Nemagraptus gracilis* lived. Hedberg (1941; 1948) and the American Commission on Stratigraphic Nomenclature (1952) tend to place biostratigraphic zones more or less on the same footing with mineral zones, or any other zones based on physical rock properties. This seems to be a fundamental mistake, for nobody has been able to trace a mineral zone around the world and suggest that it may serve as a reliable time line. Until now this has only been possible for biostratigraphic zones.

Allan (1948) believes it to be a "widely accepted view" that dissimilarity of faunal content indicates difference in age, and that similarity of faunal content establishes chronological equivalence. Such notions were already discredited by Gressly in 1838 and by many later geologists. Modern biostratigraphers will generally agree with Allan when he characterizes such statements as "extremely unsafe generalizations". Few paleontologists now entertain the idea that mere similarity of faunal content establishes chronological equivalence. Strict time equivalence, on the geological scale, is indicated by identical faunas. According to circumstances "identity" may be defined on the specific, generic, or even familial level. No general rules can be laid down in this respect, because criteria used in differentiating taxonomic units in different phyla, or even classes and orders within the same phylum, are not always commensurable.

Some modern critics of the biostratigraphic method seem to have fallen into a logical trap. It is well known that from the sentence "all men are vertebrates", it does not follow that "all vertebrates are men". Similarly, if faunas and floras may be zone-equivalent, yet not alike, it does not follow that faunas and floras may be alike, yet not zone-equivalent. Consequently, we also reject the "paratime-unit" concept of Wheeler *et al.* (1950), together with their statement that "biostratigraphic units are intrinsically transgressive with respect to time". If this were true, no earth history could have been written. If wisely chosen, properly defined, and well-tested biostratigraphic units are "para-time-units", then all our stratigraphy is parastratigraphy and historical geology parascience.

Minor Biostratigraphic Units

Although the zone is the most important and fundamental unit in biostratigraphy, others may be useful in more specialized ways. The confusion that exists in the interpretation of the zone concept extends to other biostratigraphical concepts and terms. In the following discussion I shall attempt to distinguish between rocks and time, that is, between biostratigraphical and biochronological concepts. Schindewolf (1950a, p. 28) listed 38 terms that have been used for biostratigraphical and biochronological units. His list is not complete, yet I agree with Schindewolf that most of those terms are of little or no practical value.[4] The

[4] The new term "assemblage zone" recently pro-

following discussion is limited to a selection of concepts which have received—rightly or wrongly—widespread use in the literature or seem significant.

Since the time equivalent of a zone is about the shortest time interval which can be discerned universally in the geological past, it follows that subdivisions of a zone can be of local interest only. Subzones have long been recognized and used to indicate finer biostratigraphic subdivisions where the fossil record is suitable and abundant. Although the maximum range of a subzone will as a rule be of the order of magnitude of a continent, there are exceptions. Thus, subzone alpha of the Upper Devonian *Manticoceras* zone, which can be followed over much of Europe and into North Africa, has also been recognized in northwestern Australia (Teichert, 1943) and on the North American continent (Miller, 1938). Such a "subzone", however, is equivalent to a zone in some other geological systems, and redefinition of biostratigraphic terminology seems here indicated.

"Faunizone" was originally proposed by Buckman (1902) for "belts of strata, each of which is characterized by an assemblage of organic remains". This definition is the same as Oppel's definition of a zone, and the two terms are synonymous. Usage has long sanctioned the use of faunizone for biostratigraphic zones characterized by animal fossils. The corresponding florizone, then, is a zone distinguished by plant fossils. Thus defined both terms are useful and unambiguous.

"Zonule" was proposed by Fenton and Fenton (1928) and is best used in the more concise definition of Kleinpell (1938, p. 90) for "locally recognizable biostratigraphic units"; it is therefore generally below the level of a zone and rarely contributes to the orthochronologic time scale. Thus, we may define an *Amphipora* zonule in the Upper Devonian of Arizona or even the Rocky Mountain area, where *Amphipora* may have a time significance which it does not possess outside this geological province.

In conclusion, a zonule is a biostratigraphic unit that is recognizable in a sedimentary basin or similar restricted area of sedimentation; a subzone is recognizable across a continent; and a zone is recognizable beyond continent limits, preferably world-wide.

posed by the American Commission on Stratigraphic Nomenclature (1957) (= cenozone of Moore, 1957) for what in this paper is called biostratigraphic zone has been discussed elsewhere (Teichert, 1957).

Time Concepts in Biostratigraphy and Biochronologic Units

In discussions concerning biostratigraphic units above the zone level time and rock concepts have become inextricably mixed and confused. The former have achieved generally more attention and will therefore be discussed first.

For those who uphold the spatial or rock concept of the zone, it is a matter of convenience, almost of necessity, to adopt a term for the time equivalent of a zone, which, as has been pointed out, is probably the shortest perceptible time interval in earth history. Definition of such a biochronologic unit is desirable in order to cover time intervals for which more than one zone has been defined on different paleontologic evidence in different areas of sedimentation.

The need for this was felt early, and the first International Geological Congresses gave much consideration to questions of time and rock terminology at the stage and zone level. The matter was first discussed at the Congress in Bologna in 1881, where the only concrete suggestion came from the Swiss delegation (Renevier, 1882), which formally proposed "moment" for the time equivalent of a zone. The term "phase" seems to have been discussed at this Congress but was not proposed formally until the Paris congress in 1900 (Renevier, 1901). Therefore the term "moment" has priority, as was also recognized at an early date by Diener (1919). Later, Jukes-Brown (1903) suggested "secule" for the same thing, and more recently, Kobayashi (1944) proposed "zone-time" and "instant". Zone time is suggested as the average duration in years of a zone in any given geological system and may range from system to system between about 300,000 and 5 million years. An "instant" is, somewhat cryptically, defined by Kobayashi (1944, p. 742) as a "stationary point in time-current". Kobayashi seems to want to use it for the time corresponding to the deposition of a zone or subzone as used in this paper. However, since "moment" and other terms are already available for such time units, the term "instant" may suitably be retained for a time interval of no appreciable length. Thus used, it is a somewhat theoretical concept, representing the beginning or the end of a biochronologic unit. In the terminology of some authors an "instant" may be thought of as a true "time plane". This may not necessarily be world-wide.

Although the demand for designation of a biochronological equivalent to the biostratigraphic zone seems to be obvious, the prevailing confusion can probably be resolved only by international agreement. The diverse uses to which the word "phase" is put in various geological sciences seem to make it less useful, and "moment" also seems to have the merit of priority. In spite of Arkell's (1933, p. 21) objections a term like moment recommends itself for the time equivalent of a zone because it is the shortest recognizable time unit into which, for all practical purposes, geologic time can be subdivided.

Concepts to be considered next are "biochron", "biozone", and "hemera". Biochron was proposed by H. S. Williams (1901, p. 579–580) as a time term for the "absolute duration of a fauna or flora or component parts of it". Thus the biochron of the trilobites would be the Paleozoic era, that of *Globotruncana* the Late Cretaceous. In spite of its obvious usefulness and practicality the term never "caught on". Arkell (1933, p. 22) attempted to revive it but has had few followers. In its stead the ambiguous concept of biozone became established in the literature.

Biozone was proposed by Buckman (1902) "to signify the range of organisms in time as indicated by their entombment in the strata". As thus first defined the term is an objective synonym of Williams' biochron, but since at that time biostratigraphic and biochronological thinking was more active in Europe than in North America, Buckman's biozone found acceptance in European literature from which it was later transplanted to North America where it usurped the place of the indigenous "biochron". It would be unrewarding to attempt to follow the vagaries of the term biozone through European and American literature. German writers also applied the word to rocks deposited during the time interval for which it had been coined. A reversal of the original meanings of the terms zone and biozone thus took place.

In America, most writers retained the time connotation for biozone which Buckman had given it, and the word still is widely used to denote the time span of taxons of any category, although Moore (1948) found the concept "to have little practical use". In view of divergent usage of the term, however, Arkell's (1933) proposal to revive Williams' biochron is to be strongly recommended, apart from the fact that it has priority. The term biozone should then be dropped. As a chronological term it is preceded by biochron; as a stratigraphic term it is useless. For small units we have species and generic zones, faunizones, and florizones; for rocks deposited during the survival time of a family, order, or phylum, no stratigraphic term is required. The term "range zone" recently proposed for stratigraphic units of this type by the American Commission on Stratigraphic Nomenclature (1957), or "acro-zone" of Moore (1957), cannot be considered a biostratigraphic term (Teichert, 1957).

Hemera is another biochronological concept of doubtful merit. The term was proposed by Buckman (1893) with a definition which is not very clear but has generally been interpreted to refer to the acme of development of one or more species, thus not covering the entire time span of such species. Trueman (1923) proposed "epibole" as a "stratigraphical term to cover deposits accumulated during a hemera." Since the acme of the development of a species commonly coincides with the time during which it is characteristic of a zone, there is obviously considerable overlap between epibole and zone. Also the acme of development of a species is in many cases confined to a certain area. Hence, epibole and hemera should probably not occupy places in a formal biostratigraphical or biochronological nomenclature. More and more they seem to have approached the meaning of zone and its time equivalent.

Finally, we must briefly consider the peculiar concept of "Teilzone" as it developed in German literature. Pompeckj (1914) introduced the term for part of a zone (which he regarded as a time unit) to designate the "local duration of existence of a species". This obviously makes it a locally recognizable part of a Buckman biozone and thus a time, or biochronological, term. The term was thus widely used in German literature. Arkell (1933) analyzed the concept but seemed to think that the Germans generally applied it in a spatial sense. For this Moore (1957) proposed the term "topozone" which will probably be found useful to designate paleontologic zones recognizable in a single locality. As a time term corresponding to the supposedly spatial, or biostratigraphic, teilzone (= Moore's topozone) Arkell proposed "teilchron", but this term is synonymous with the original Pompeckj Teilzone. Since the local duration of a species can only be determined by observation of its distribution within a zone, and the local duration of individual species during the deposition of a zone may

be controlled by facies and thus be without time significance, an analysis of "teilzones" as time units contributes little to the biochronologic standard.

In conclusion, I believe with Arkell and Schindewolf that only a minimum number of biostratigraphic and biochronologic terms is required. The concepts and terms discussed above are listed below. Capital letters indicate terms that are recommended, ordinary print terms toward which I am indifferent; and terms in parentheses are regarded either as synonyms or as terms whose use is not recommended.

Biostratigraphic Units	Biochronologic Units
ZONE, d'Orbigny, 1850; Oppel, 1856–1858 (*Synonyms:* faunizone, Buckman, 1902; Orthozone, Kobayashi, 1944; assemblage zone, American Commission on Stratigraphic Nomenclature, 1957; cenozone, Moore, 1957) A bed or beds characterized by one or more distinctive fossil species or genera, used for intercontinental to world-wide correlation.	MOMENT, Swiss Subcomm., Bologna 1881, *see* Renevier, 1882; Diener, 1919 (*Synonyms:* phase, Int. Cong. Paris, 1900, *see* Renevier, 1901; secule, Jukes-Browne, 1903; hemera, Buckman, 1902, not 1893; Zone, Waagen, 1869, and subsequent German authors) Time during which a zone was deposited.
SUBZONE, of authors Subdivision of zone, rarely of world-wide importance. ZONULE, Fenton and Fenton, 1928; Kleinpell, 1938 Minor biostratigraphic units recognizable in geographically restricted areas.	(No time terminology required)
(Biozone, German authors and Arkell, 1933; range zone, American Commission on Stratigraphic Nomenclature, 1957; acrozone, Moore, 1957) Rocks deposited during a biochron.	BIOCHRON, Williams, 1901 (*Synonym:* biozone, Buckman, 1902) Total time of existence of any taxonomic unit (species, genus, or higher).
TOPOZONE, Moore, 1957 (*Synonyms:* Teilzone, some German authors and Arkell, 1933; zone in terminology of many authors) Paleontologically defined horizon recognizable in a single locality.	(Teilzone, Pompeckj, 1914) (*Synonym:* Teilchron, Arkell, 1933) Locally recognizable time of existence of a taxonomic unit
Epibole, Trueman, 1923 Rocks deposited during a hemera.	Hemera, Buckman, 1893 Time of the acme of development of a taxonomic unit, especially species.
(No corresponding biostratigraphic unit)	INSTANT, Kobayashi, 1944 A true "time plane," the beginning or the end of a moment, or other biochronologic unit; identified by tops or bottoms of biostratigraphic units.

Biostratigraphic vs. Time-Stratigraphic Concepts

The difference between biostratigraphic and time-stratigraphic units has been briefly indicated above: time-stratigraphic units are rocks formed during a certain time interval; biostratigraphic units are rocks identified by a distinct fauna or flora. Need for clear separation between biostratigraphic and time-stratigraphic concepts is convincingly demonstrated in Arkell's (1956) succinct presentation of what may be called the mechanics of stratigraphic correlation of the Jurassic rocks of central Arabia with those of the English standard section.

Biostratigraphic units should be identified with a type section or type locality in the same way in which rock-stratigraphic units are identified. The type locality or section serves as an objective standard with which spatially separated parts of the unit may be compared. Time-stratigraphic units have no "type sections". This is not generally understood, although obviously any stratum or stratigraphic section deposited during an arbitrarily chosen time interval is typical for that interval because it was formed during that interval. The search for a "type Permian", a "type Triassic", or any other type units in time stratigraphy has consumed much time and effort over the years but has brought no tangible results, because they do not exist.

Typical Permian or Triassic exists wherever Permian or Triassic fossils are found. However, in view of the kaleidoscopic variety of organisms and environments that existed side by side all over the world at any given time, selection of any one group of organisms or of any one organism-environment combination as "the" typical one for that time can hardly be meaningful.

The only rock units that may be called typical for any particular time are its recognizable biostratigraphic units, because every such unit contains a unique fossil assemblage. The only practical time-stratigraphic boundaries are interfaces between two superimposed biostratigraphic units, selected in an area where the succession of biostratigraphic units during the critical interval is as complete as possible and from which these units can be traced to other parts of the world. Thus, the Ordovician period can be defined as the time interval between the beginning of the deposition of the zone of *Dictyonema flabelliforme* and the end of deposition of the zone of *Dicellograptus anceps*. All and any rocks deposited during this time interval anywhere in the world are typical for the Ordovician period. They constitute the Ordovician System. Where the diagnostic zone faunas, or proved equivalents, are not represented in the critical part of a section, boundary problems usually arise.

Whereas a time-stratigraphic unit can have no single type section or locality, it may have a number of standard sections which exemplify the unit in geographically restricted areas (Dunbar and Rodgers, 1957, p. 301). Thus, there is a standard section for the Triassic System in the eastern Alps, another one for the Triassic System in Germany, and yet another one for the Colorado Plateau. The first has little in common with the second, and nothing with the third.

Conclusion

In an era of glorification and triumph of the physical sciences we are sometimes tempted to believe that all geological issues may be amenable to analysis and solution by experiment and mathematical formula and to forget that geology occupies the borderland between the physical and the biological sciences. Geology emerged as a modern science only when it had been discovered that the study of fossils made it possible to sort out the chaos of rocks on the surface of the earth and to arrive at a reasonably logical historical interpretation of geologic events during a considerable part of geologic time. This basic situation still exists in spite of impressive advances in the study of physical properties of rocks. Representation of geologic data by formulas and diagrams in many cases conceals lack of accurate knowledge and simulates a degree of accuracy to which geology cannot yet aspire. Also, geologic results achieved by application of physical methods are only as good as these methods, and errors of measurements may be greatly magnified by the geologic time factor involved.

Inasmuch as science, by definition, should be able to predict situations or events, some physicists have counted geology and paleontology among the historical sciences which "predict backward". (*See* Schrödinger, 1956, p. 192–193.) However, predictions by geologists of mineral deposits, oil pools, or succession of fossil faunas are not different in nature from, for example, the prediction of unknown subatomic particles by nuclear physicists. Both predict the presence of something which has been in existence but has escaped recognition. Some predictions made by biostratigraphical methods are of this kind.

Measurements of past time periods depend on observational methods basically similar tc those applied in the measurement of contemporaneous time. They are deductions from observations of rock properties or structures (including fossils) which may be interpreted as having resulted from the operation of processes which are either unidirectional and irreversible or occur with a fixed periodicity. The only terrestrial phenomena which fulfill these conditions and have proceeded universally for considerable periods are radioactive disintegration and the evolution of life. Paleontologic studies supply an accurate and dependable space-time system of reference for which radioactive dating furnishes an approximate scale in terms of absolute time. Within the wide frame of reference supplied by a certain number of radioactive dates, dating by paleontologic methods (biochronology) is considerably more accurate than dating by any known physical method, except perhaps for the last few thousand years of earth history.

In this paper, stratigraphy has been defined broadly as the geologic science that is concerned with the study of rock strata for the purpose of establishing their chronological order and their geographic distribution. Biostratigraphy is the approach to stratigraphy through paleontologic methods. Its purpose is the recognition and definition of biostrati-

graphic units in sedimentary sequences. A biostratigraphic unit is a bed or series of beds that is characterized by one or more distinct fossil species, which serve to differentiate it from other biostratigraphic units. The fundamental biostratigraphic unit is the zone.

Ordinarily the zone is the smallest biostratigraphic unit for which world-wide contemporaneity of deposition may be objectively established, although not all zones have a world-wide distribution. Zones can be recognized only on the presence of their defined zone fossils.

Corresponding to biostratigraphic units are biochronologic units. A biochronologic unit is the unit of time during which sedimentation of a biostratigraphic unit took place. Biochronologic units are true relative time units. They are based on objective paleontologic data. The biochronologic unit equivalent to the zone, for which the term moment is preferred, is for all practical purposes the smallest time interval for which geographical and biogeographical conditions of the past can be summarily described. This time interval varies probably between about half a million and 5 million years.

Time-stratigraphic units are units of rocks which have been deposited during a defined unit of time. The upper and lower limits of such time units must be more or less arbitrarily determined by biostratigraphic methods. Arrangement of rock-stratigraphic units within a time-stratigraphic unit and assignment to such a unit are generally made on a variety of lines of evidence both physical and paleontologic, including extrapolations, conclusion by analogy, and sometimes merely for reasons of expediency. Accuracy of correlation of a time-stratigraphic unit and its various parts increases with the number of biostratigraphic units, especially zones, that it contains.

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Short Notes

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SIGNIFICANCE OF COCCOLITHOPHORIDS IN CALCIUM-CARBONATE DEPOSITION

BY M. N. BRAMLETTE

Investigation of the living coccolithophorids by marine biologists is receiving increased attention as these forms constitute an important element of the oceanic plankton. Among the more important investigations are those of Braarud and associates at Oslo, Kamptner and predecessors at Vienna, and Deflandre in Paris. Even a brief review of present knowledge is not attempted in this paper, as this has been presented recently (Bramlette and Riedel, 1954, p. 385–386) and much more thoroughly considered by others (Deflandre, 1952, p. 439–470).

Coccolithophorids are minute beflagellate Protista that derive their food supply through photosynthesis and are usually classed as algae. The skeletal material consists of a number of coccoliths (about 2–20 microns) at or near the surface of a more or less globular body of protoplasm. The coccoliths are usually not well enough articulated to remain together in the sediments as complete skeletal remains (coccospheres), though these are found in some samples. X-ray-diffraction studies indicate that they consist of calcite in the bottom sediments, and if any originally form as aragonite these must be quickly dissolved or converted to calcite.

Studies on the stratigraphic distribution of coccolithophorids include preliminary examination of more than 1000 samples and indicate much promise for these neglected fossils in correlation and age assignments. The wide distribution of these planktonic organisms is strikingly evident in the many identical forms in strata assigned on other evidence to the middle Eocene of France, Louisiana, and California. Many samples must be examined before stratigraphic ranges are reasonably well established, but it now appears that their vertical ranges are comparable with other fossils, such as Foraminifera. For example, a peculiar and easily identified form, described by Deflandre from the upper Eocene of New Zealand, has now been found in the upper Eocene only of such widely separated regions as Australia, Trinidad, Alabama, and France.

Samples from the type Danian (generally considered uppermost Cretaceous) show an assemblage very different from that of the underlying Maestrichtian, and a closely comparable assemblage is found in the Midway of Alabama (usually assigned to the Paleocene) overlying an assemblage like that of the Maestrichtian. Though some discontinuity is evident between the type Danian and underlying Cretaceous, it does not seem to mark any very great time break, and the marked difference in the assemblages may prove to be largely due to some ecological difference such as change in surface-water temperatures. The coccoliths of both the Danian and Midway are more like those of the early Tertiary than like those of the underlying Cretaceous. This and other lines of evidence remain to be considered thoroughly, however, before an age equivalence between the Danian and Midway might be suggested.

A number of distinctive coccolithophorids seem to have developed near the end of the Oligocene or in the early Miocene, as interpreted from the sequences in Italy, Trinidad, and Indonesia. Some additional evidence may thus be available on the Aquitanian and equivalents elsewhere, including the E and F part of the lettered sequence used for the Tertiary of Indonesia.

The earliest known occurrence of well-preserved and definitely recognizable remains of coccolithophorids is in the Jurassic (Deflandre, 1952, p. 459–460), but they occur in great vari-

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TABLE 1.—RELATIVE IMPORTANCE OF COCCOLITHS, FORAMINIFERAL TESTS, AND OTHER CONSTITUENTS IN CALCAREOUS OOZE AND CHALK

(In per cent of total sample)

	Recent Globigerina ooze of deep Pacific Albatross 62 (56-60 cm)	Miocene calcareous ooze of deep Pacific Chubasco 24 (20-24 cm)	Oligocene (upper?) coccolith ooze of deep Pacific Chubasco 38 (10-12 cm)	Oligocene (middle?) coccolith ooze of deep Pacific Midpac 40 (130-140 cm)	Oligocene (lower?) coccolith ooze of deep Pacific Midpac 5-1 (50-60 cm)	Eocene (middle?) coccolith ooze of deep Pacific Midpac 25E (50-60 cm)	Cretaceous (Maestrichtian) chalk from Stevns Klint, Denmark	Cretaceous (Campanian) chalk from Travis County, Texas
Total calcium carbonate (from chemical analysis)	86	79	87	91	84	95	93	65
Coccoliths (including discoasters and related forms)	10	26	35	58	48	43	18	21
Foraminifera (including identifiable fragments)	55	25	30	5	15	32	23*	18*
Other calcareous skeletal debris	3	8	5	1	3	5	15	10
Indeterminate calcite particles	18	20	17	27	18	15	37	16
Radiolaria and diatoms	9	13	9	0	7	0	0	0
Inorganic non-calcareous residue	5	8	4	9	9	5	7	35

* In large part benthonic forms

ety and vast numbers in the Cretaceous chalk formations. They form an even greater part of the calcareous deep-sea deposits of the Tertiary, though they are of less importance in the late Tertiary and constitute only a minor part of most Recent calcareous ooze.

Table 1, showing the results of a careful examination of eight samples of calcareous pelagic deposits that appear to be fairly representative of many others examined less thoroughly, indicates the relative importance of the coccolithophorids, Foraminifera, and other significant constituents. An accurate determination of the percentages for a sample does not seem practicable, as any separation of the very small calcareous constituents is impossible. Most of the Foraminifera and siliceous shells are large enough for separation and fairly accurate weight determination. The indicated percentages of Foraminifera are nearly 100 per cent too high in the Cretaceous samples, however, because their chambers are filled with secondary calcite. A count of the finer constituents within fields under high magnification can give only approximations, as a size separation was unsatisfactory, and size ranges between about 1 and 20 microns had to be considered. Enough fields were examined, however, to indicate relative proportions. The accuracy of Table 1 is not determinable because of the great size differences. The percentages might, at least, be rounded off to the nearest 5 per cent. The percentages for the major constituent are even less precise, as they were adjusted so that the total of calcareous constituents agrees with the chemical determination of total calcium carbonate.

The distinctive interference patterns of coccoliths observed under crossed nicols (Kamptner, 1954, p. 1–51) are a considerable aid in determining their number in a field under the polarizing microscope. The percentages of indeterminate calcite particle doubtless include small fragments of the recognized groups and perhaps usually in about their same relative proportions. This is not the case, however, in the chalk sample from Denmark. As the later discussion indicates, this and many other Cretaceous chalk samples may consist largely of crystals of secondary calcite. Among noncalcareous skeletal components, only the Radiolaria and diatoms (together) are tabulated, and the "residue" consists largely of clay and other detrital minerals.

The Cretaceous chalk formations contain abundant coccoliths, though these are commonly not the dominant calcareous constituent and are not in the two samples in Table 1. No deep-sea pelagic deposits of Cretaceous age are now available, however, and these two samples represent relatively shallow epicontinental sea deposits with many benthonic Foraminifera and other fossil remains. One might expect that much of the tiny remains of coccolithophorids would be swept out into deeper water. Their abundance in the Cretaceous chalks as compared to deposits formed under comparable conditions in the Tertiary—when the coccoliths dominated the calcareous pelagic deposits—suggests that the coccolithophorids were at least as dominant in the calcareous plankton of the Cretaceous. This would be more strongly indicated if the loss of smaller coccoliths through dissolution and recrystallization is as important in the Cretaceous chalks as the following discussion suggests.

Some samples of Cretaceous chalk, such as one from the Selma Formation of Mississippi, obviously contain a greater percentage of coccoliths than do the two considered more or less representative. Some others, however, contain lower percentages of coccoliths, particularly those examined from England and Germany. These samples show much corrosion and secondary growth of the coccoliths. Their content of small calcite particles, higher than the 37 per cent tabulated for the Maestrichtian sample of Denmark, may represent a corresponding loss of coccoliths through recrystallization, with dissolution apparently accounting for the comparative sparsity of the normally abundant smallest coccoliths. The induration is generally greater in such samples, and commonly the chambers of Foraminifera have been filled with calcite. Less corrosion and recrystallization are usual in less permeable clayey chalks.

The many cores of Tertiary calcareous sediment obtained from the deep Pacific may be classed as coccolith ooze. Those of middle and early Tertiary age, particularly, show a consistent predominance in total mass of coccoliths over Foraminifera (largely planktonic), though in those of late Tertiary age this is less marked. In the post-Tertiary, however, Foraminifera (largely planktonic) greatly predominate. These more recent deposits have thus been commonly called *Globigerina* ooze, though this term is not entirely appropriate as the genus *Globigerina* does not everywhere predominate in pelagic foraminiferal ooze.

In much of the Atlantic Ocean, the percentage of coccoliths in the modern deposits seems to average even less than the 10 per cent

35 7 5 9 9 4 8 5

calcareous residue

* In large part benthonic forms

shown in the recent sample from the Pacific (Table 1), but their dominance in the Tertiary seems to be comparable with that in the Pacific. Some cores of Tertiary age, collected from the Atlantic by Lamont Geological Observatory, were described (Ericson, Ewing, and Heezen, 1952, p. 504–507) as fine-grained chalk, and some of these samples examined by the writer contained a great abundance of coccoliths, though the samples were so small that relative percentages were not determined.

There is evidence that, during much of Tertiary time, ocean waters were warmer than at present, and oxygen-isotope data of Emiliani and Edwards (1953, p. 887) indicate that temperatures of the bottom waters may have averaged about 12° C. This would have lowered the solubility limits of calcareous deposits. The carbonate is dissolved, and little accumulates now below depths of about 4500 m in the Pacific Ocean and below somewhat greater depths in the Atlantic Ocean. Revelle has calculated (personal communication) from his graphs of the controlling variables (Revelle and Fairbridge, 1957) that the present critical depth of about 4500 m would correspond to about 6700 m for comparable solubility conditions with a Tertiary bottom-water temperature of 12°C. This would explain the occurrence of Tertiary calcareous ooze beneath "red clay" in some cores from areas where depths are too great for calcareous accumulation under present conditions. It does not seem that this solubility difference for different bottom temperatures could account for the relative proportions of coccoliths and Foraminifera in the Recent and Tertiary calcareous sediments, however, because even the smallest coccoliths are present in both, and they show little more susceptibility to dissolution than do the Foraminifera—as judged by their distribution with depth on present ocean bottoms. This is in peculiar contrast to the distinctly greater susceptibility of coccoliths to solution and recrystallization in the intrastratal waters of Cretaceous chalk formations.

The coccolithophorids thus have had a significant role in the geochemical cycle and balance of calcium carbonate in the ocean basins. An interesting and often-quoted suggestion by Twenhofel (1939, p. 327), later elaborated by Kuenen (1950, p. 392–394), was that *Globigerina* and other planktonic Foraminifera, which first appeared during the Cretaceous, have since greatly modified the major site of deposition of calcium carbonate and hence its whole geochemical cycle in sediments. The great amount of modern *Globigerina* ooze in the deep-ocean basins represents a loss to this cycle, in contrast with the great and presumed dominance of accumulation in shallow (perhaps epicontinental) seas of pre-Cretaceous time. Much of this earlier deposition in epicontinental or marginal seas must have been incorporated again into the continental masses and returned to the cycles of erosion and deposition. The large loss to the deep seas under modern conditions was calculated by Kuenen to indicate a potential calcium-carbonate "famine" in 100-150 million years. The lesser amount postulated as lost to the deep seas in earlier times would help account for the high proportion of limestones among older rocks of the continental areas.

Kuenen (1950, p. 393) also suggests that "the present relative scarcity of calcareous deposits in shallow waters, as compared to former periods," may be similarly explained. This scarcity seems questionable in island-atoll areas, however, and in shelf-sea areas seems more probably to reflect only a greater dilution with clastic sediments from rapid erosion of unusually great topographic relief in late geological time.

It is not yet apparent that the earlier dominance of coccolithophorids was due to a greater rate of production than at present, nor that there was a subsequent (post-middle Tertiary) increase in the rate of production of the pelagic Foraminifera. The latter possibility would help the budget problem of calcium carbonate in the oceans and seems to fit the meager data suggesting a greater-than-average rate of deep-sea deposition in Quaternary time. It is not too evident, however, that the rate of supply and consequent deposition of calcium carbonate has varied greatly, and perhaps only the relative importance of the two groups of calcareous plankton has changed. At least it seems evident that the rate of supply is the critical factor for much change in the rate of carbonate deposition. At present the oceans are about saturated with calcium carbonate, and the supply has been calculated to be sufficient to replace this total amount within roughly a million years (Revelle and Fairbridge, In press).

Though the coccolithophorids seem to have been more important than the pelagic Foraminifera before late Tertiary time, their importance is not demonstrable before the Cretaceous. Unless the coccolithophorids were important earlier, however, their dominance over the

pelagic Foraminifera would not alter the significant conclusion of the hypothesis on calcium-carbonate loss to the deep-sea bottoms through pelagic accumulations since Cretaceous time. Kuenen's suggestion seemed particularly inviting because both the geochemical budget during later geologic time and the excess of calcium carbonate in known pre-Cretaceous rocks present a less serious problem if the present large loss to the deep sea is abnormal and most of the carbonate was deposited on or near continental blocks during much of geological time. Attractive as this hypothesis appears, it would have no basis if such organisms as coccolithophorids or an inorganic process could have segregated much calcium carbonate in the deep-ocean bottoms before the advent of pelagic Foraminifera; both possibilities seem to need consideration.

Very fine-grained limestones in especially thin sections under a high-power microscope show a fine mosaic texture similar to aplitic—or xenomorphic granular—texture in igneous rocks. The small (commonly only a few microns) interlocking crystals indicate a recrystallization of whatever small particles were originally present. One occurrence where these were evidently coccoliths will be described. Some may equally well have been like the tiny needles of aragonite now forming on the Bahama Banks, which have been called drewite. This drewite is considered here as probably of dominantly inorganic or at least nonskeletal origin. Though this has been questioned (Lowenstam, 1955, p. 270–272), its origin is not critical to this discussion. Such material has not been found in careful examination of many Recent and Tertiary deep-sea deposits. Drewite deposition would seem more probable in warm, shallow waters, and more certainly so if it should prove to be disintegrated debris of algae such as *Halimeda*, as suggested by Lowenstam.

The fine-grained Carmel limestone of Utah might well represent a deposit comparable to the drewite of the Bahama Banks. It is comparatively thin, occurs between nonmarine formations, and thus was probably formed in relatively shallow water. Thin sections of some parts also show small pellet forms like the oölites of the Bahama Banks deposits.

Very fine-grained limestones are common, however, and rather characteristic of some geosynclinal formations that seem to represent relatively deep-water accumulations. The Tithonian and other limestones of the sub-Alpine region of France are examples.

The requirements for inorganic precipitation of much calcium carbonate include warming of the water, or decrease of carbon dioxide, or increase of salinity through evaporation. In a former time without calcareous planktonic organisms, one or more of these requirements might have been maintained to a sufficient extent near the surface of relatively deep waters, despite some inevitable mixing with water from below the surface, and inorganic precipitation might have occurred here as well as in shallow water. Conceivably the fine-grained limestones of geosynclinal accumulations could have originated in this way. Assuming inorganic precipitation under these conditions, however, would make it equally possible in the open ocean and offers no support to the hypothesis of a change to a greater proportion of deep-ocean accumulation of calcium carbonate during Cretaceous and later time.

This hypothesis is not supported by evidence of greatly increased importance of calcareous plankton either unless forms such as the coccolithophorids were, like the pelagic Foraminifera, insignificant before about this time. The coccoliths actually are found in great abundance only with the earliest occurrences of chalk in the Cretaceous, but the following considerations suggest that earlier occurrences may be represented by some pre-Cretaceous fine-grained limestones. Thus the problem would remain regarding a geochemical balance for calcium carbonate.

The original absence or paucity of planktonic Foraminifera is evident in pre-Cretaceous limestones which show comparably robust microfossils preserved, despite the evident recrystallization of the much finer matrix particles inherent in lithification to a dense limestone. Though not recognizable, the original presence of the minute coccoliths is indicated in a Cretaceous limestone of Tunisia, and this example suggests that their preservation is not expectable in other fine-grained limestones, and especially in those of greater age.

The upper Cretaceous chalk formations in central Tunisia are represented by fine-grained limestones in the thick and folded geosynclinal sequence to the north. A thin section of the Maestrichtian part of the limestones shows Foraminifera (mostly pelagic) scattered in a matrix of interlocking calcite crystals which average a few microns—just as these Foraminifera are scattered in the abundant coccoliths of the less-deformed chalk equivalent to the south.

The thick section of geosynclincal deposits of the sub-Alpine region of France includes much very fine-grained limestone of Jurassic and Cretaceous age. Thin sections show that the interlocking grains of the dense Tithonian (upper Jurassic) limestone average little larger than the sizes of coccoliths which occur in a variety of forms in the somewhat older, though much less lithified, Oxfordian marl of northern France. The only common fossil recognizable in this Tithonian is a small form classed with the planktonic tintinnids. The skeleton of this *Calpionella* is considerably larger and more robust, and thus more apt to be preserved, than any of the coccolithophorids.

This little evidence obviously does not justify any generalizations about fine-grained limestones, but the possibility that geosynclinal ones, at least, were originally like the later coccolith chalks seems to offer fewer difficulties than does an inorganic (nonskeletal) origin. This would also explain an otherwise unrelated coincidence in the occurrence of coccoliths in abundance and great variety of forms along with the earliest occurrences of the relatively unlithified calcareous deposits known as chalk. The absence of the coccoliths and their progenitors in the older fine-grained limestones would be inevitable with the recrystallization that has evidently produced the aplitic texture and obliterated whatever form the small particles originally showed.

These considerations seem to indicate, at least, that no tacit assumption that these limestones represent a fine inorganic precipitate is justified, merely because no very small fossil remains are still recognizable. Geochemical studies may help the problem for any particular limestone, as the distribution of such minor elements as barium and strontium in inorganic and skeletal calcium carbonate becomes better understood; and the carbon and oxygen isotopes may offer some evidence despite the recrystallization.

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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LOWER CAMBRIAN SECTION OF VERMONT

BY CHRISTINA LOCHMAN

W. M. Cady has called the writer's attention to a few corrections which should be made in the diagram of the Lower Cambrian succession of northwestern Vermont and adjoining southern Quebec (Lochman, 1956, p. 1353, Fig. 1) in order to bring it in line with the most recent observations and interpretations. The corrected and supplemented diagram is presented (Fig. 1).

The formational boundary between the Dunham dolomite and the overlying Winooski dolomite should be placed at the position of maximum areal expanse of the Monkton and Parker clastic facies. Cady considers some of the lower Winooski as probably of Early Cambrian age, particularly where it grades laterally into strata of the upper Monkton facies. Shaw (in press) in the Saint Albans area finds a regional unconformity beneath the Parker, between it and the underlying Dunham dolomite, and also an unconformity at the top of the Parker, between it and the overlying Rugg Brook formation. The position of these unconformities is shown on the diagram by the heavy wavy line. Still farther east Cady has observed no positive evidence of unconformity, and the unconformable contacts are shown passing into conformable relations; he finds the underlying Dunham dolomite and the Middle Cambrian Rugg Brook strata grading laterally into a lithofacies like that of the Parker but probably younger. The eastern

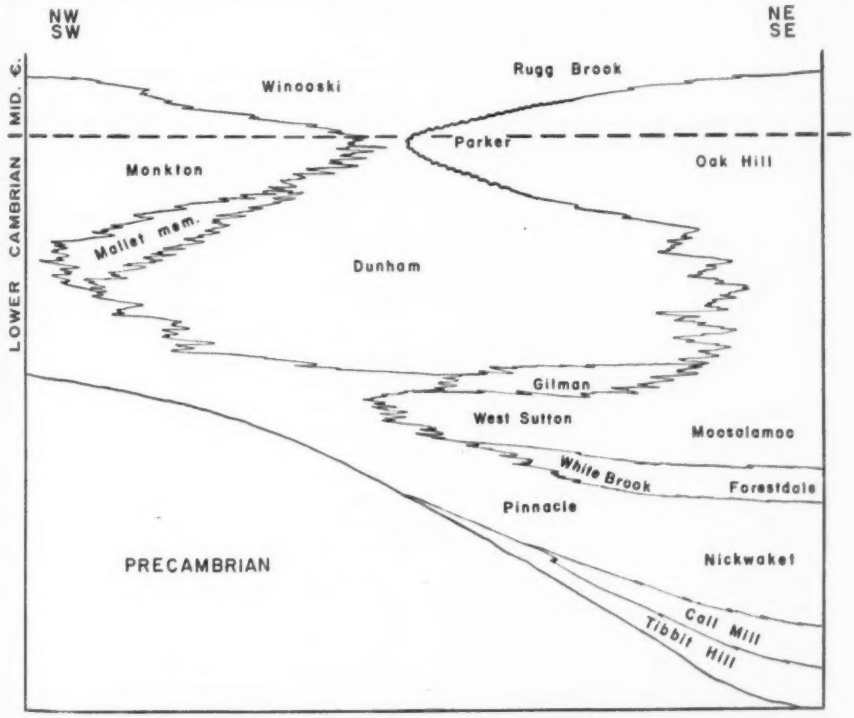

FIGURE 1.—CORRELATION OF LOWER CAMBRIAN FORMATION OF NORTHWESTERN VERMONT AND ADJOINING SOUTHERN QUEBEC

area was probably a belt of subsidence in which evidence of episodes of nondeposition or erosion would be theoretically lacking or negligible and unconformities less likely than closer to the platform as in the Saint Albans area. In Figure 1 Tibbit Hill, Call Mill, White Brook, West Sutton, Gilman, Dunham, and Oak Hill are unit names applicable to northwestern Vermont and adjacent southern Quebec. Nickwaket, Forestdale, and Moosalamoo are names applicable in west-central and southwestern Vermont.

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OLIVINE-SPINEL TRANSITION IN FAYALITE

BY A. E. RINGWOOD

During a study of the olivine-spinel transition and its bearing on the constitution of the mantle, an investigation was made of the conditions required for this transition in fayalite. Sub-solidus equilibria in the system Fe_2GeO_4–Fe_2SiO_4 were determined using a "squeezer" type apparatus (Griggs and Kennedy, 1956) at 700°C. and under a pressure of 25,000 bars. Ferrous orthogermanate, which possesses a spinel structure, dissolves about 25 per cent of Fe_2SiO_4 under these conditions. The fayalite in equilibrium with the spinel dissolved about 5 per cent of Fe_2GeO_4. Taking this data, and taking the increase in density for the olivine-spinel transition as 11 per cent, it is possible for one to calculate the pressure at which fayalite should invert from an olivine to a spinel structure (Ringwood, 1956; in press). The calculation indicates that a pressure of approximately 45,000 bars at 700°C. should cause the transition. It is to be expected that at lower temperatures the transition would occur at lower pressures.

This result suggested that the temperature-pressure conditions required for the transition might be in the range of the "squeezer," and accordingly an attempt was made to obtain the transition directly.

Ferric oxide, metallic iron powder, and silicic acid were intimately mixed in the proportions required to give fayalite plus a 10 per cent excess of metallic iron. The samples were held in the squeezer at 400°C. at pressures of 20, 30, 38, 50, and 60 kilobars for periods ranging between 4 and 16 hours. The resultant products were then examined by x-ray and optical methods.

The runs at 20, 30, and 38 kilobars consisted substantially of fayalite, with unreacted material and metallic iron. The runs at 50 and 60 kilobars produced considerable amounts of green, transparent, isotropic crystals. Because of their small size and the presence of metallic iron and unreacted material (which ranged from 20 to 60 per cent) the refractive index was difficult to determine accurately. A value of $1.955 \pm .020$ was found. X-ray powder patterns showed that the crystals possessed a spinel structure with a lattice constant of 8.235Å. The calculated density is 4.85 gms/cc, 12 per cent higher than that of fayalite. A run at 450° C. and 55,000 bars resulted in about 80 per cent conversion to these green spinels.

It is concluded that the new phase represents a fayalite polymorph with the spinel structure, and that the transition pressure at 400° C. lies at approximately 45,000 bars. This may be compared with the pressure previously calculated for the transition. This was also 45,000 bars, but the calculation was made at 700° C. The difference between the observed and predicted transition pressure at the same temperature is probably about 15,000 bars. This is well within the combined margins of error of both methods.

The preceding results support the validity of the method (Ringwood 1956; in press) used for predicting the occurrence and stability range of high-pressure polymorphs and are of significance in understanding the constitution of the mantle.

Further work aimed at determining the slope of the fayalite transition and investigating other aspects of the olivine-spinel transition is in progress.

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COMPLEX ORIGIN OF SILTS IN THE VICINITY OF FAIRBANKS, ALASKA

BY STEPHEN TABER

Introduction

In a paper on perennially frozen ground in Alaska (Taber, 1943) a hypothesis was outlined for the complex origin of silts forming most of the surface covering of valley slopes and floors in the extensive nonglaciated parts of Alaska. The theory was further developed in a later paper (Taber, 1953) in which prevailing winds of Alaska now and during periods of maximum glaciation were briefly discussed, together with changes in glacial activity. According to the hypothesis:

"The Alaska silts have been formed and deposited in several different ways, some having had a very complex origin. They contain material produced by glaciation, volcanic activity, organic processes and weathering." Where best exposed they "have been formed chiefly by disintegration of local country rock, the products of weathering having been moved down valley slopes by soil creep and slope wash and distributed over valley floors by occasional floods" (Taber, 1943, p. 1471–1473).

In a recent paper, Péwé makes a valuable contribution by publishing a large number of mechanical analyses, but he over-simplifies the problem and misconstrues the writer's hypothesis. He states that "Most of the upland silt is considered to be wind blown, derived from glacial outwash" (Péwé, 1955, p. 699). The bulk of this so-called "upland silt" has been termed "creek-valley silt" by Taber. The theory that silts have been formed by disintegration of local country rock is construed by Péwé (1955, p. 717–718, 723) to mean that they are "residual deposits" formed from the "underlying rock," in spite of Taber's (1943, p. 1449, 1463) statements that "deep residual soils are not formed by freezing and thawing," and "Accumulation of very thick valley silts is dependent on a high frequency of the freezing cycle at the surface of the country rock when water is present, and this necessitates removal of silt from slopes before it becomes so thick as to inhibit freezing and thawing."

A re-examination of the problem has disclosed additional evidence supporting the theory of complex origin with the bulk of the silt formed by frost action on highly susceptible rocks.

Theory of Complex Origin

All silt samples examined contained admixed material except two: (1) fresh rock flour discharged by Castner Glacier; and (2) Candle Creek silt formed by frost disintegration during a single year of exposure. Plant remains were found in all the other samples. Fresh-water diatoms and sponge spicules were absent from only three of these: (1) from the steep slope on the west side of Savage River in Mt. McKinley National Park; (2) from the Nome Coastal Plain; and (3) from Ester Creek. Volcanic ash was observed in all except one from high on the valley slope southeast of Fox and one from St. Michael Island, where, oddly, a small volcano, Mt. St. Michael, has been active in Pleistocene time. The two samples showing no admixed material were coarser-textured than typical Alaska silts, and further comminution or sorting, or both, would be necessary to convert them into such silts.

Silt formed by disintegration is subjected to repeated freezing and thawing and, therefore, further comminution, during its downslope movement. Since the silt contains practically no clay to serve as a binder, it is rapidly eroded by running water when unprotected by vegetation. Supersaturation of silt near the surface, as a result of freezing and thawing, tends to move it down slope under a cover of peat and turf, leaving behind coarser fragments. In places silt may be seen flowing out from under the vegetal blanket. Turf-banked benches and similar features also furnish evidence of the movement of silt under turf (Taber, 1943, p. 1461–1463, Pl. 2, fig. 2; Pl. 4). Soil creep, mud flows, and slope wash are likewise factors in the downslope movement of silt. Rock fragments in the silt tend to accumulate at the base of valley slopes to form the "slide rock" of miners. The abrupt and widespread change from deposition of gravel to deposition of silt is probably due to change from an arid to a cold humid climate, but crustal movements seem to have caused the heavy alluviation of trunk valleys, and this extended up creek valleys and their tributaries. A well in the Tanana Valley at Fairbanks passed through

364 feet of gravel, sand, and a little clay, and geophysical tests indicate bedrock is about 60 feet below sea level (Taber, 1943, p. 1469). Bedrock may be even lower to the south under the Tanana lowland. Yukon Flats on the north side of the Yukon-Tanana upland is topographically and probably structurally similar to the Tanana lowland. Subdrainage through silt and thick gravel deposits reduced the size of streams and therefore their capacity to transport. Beaver dams which have been found buried in the silts, and probably ice dams, have helped to cause deposition. The relative importance of all these factors needs further study.

Good exposures are few, but the prevailing rocks in creek valleys, where placer mining has been conducted, can be determined from the gravel in tailing piles. The dominant formation is the Birch Creek Schist, consisting of schists, slates, phyllites, and some limestones. A few small, widely scattered intrusions of igneous rocks are present. Laboratory experiments and field observations show that the fine-grained cleavable rocks are highly susceptible to frost disintegration.

Interbedded layers of volcanic ash in valley silts are the result of distant volcanic eruptions, but they are not found at high elevations. Ash originally deposited on divides and valley slopes has been incorporated into the silts during their transportation to valley floors. This explains the small amount of ash found in most of the writer's samples.

Peat layers and the roots and stumps of old forest zones are interbedded with the valley silts, but the best explanation for the fine plant fragments found in most samples is the same as for volcanic ash—thorough mixing during transportation.

Heavily loaded glacial streams deposit silt on flood plains and as bars in braided rivers, but on the average this silt is somewhat coarser than that discharged by glaciers, since a greater proportion of the finer grains is carried farther down stream. Wind picks up its load from these deposits; the coarser grains are dropped first, often in the form of sand dunes, while the finest material is blown farthest. This is a process of sorting without appreciable comminution.

Wind-borne glacial silt is important in the Delta River Valley, 100 miles southeast of Fairbanks, where Taber (1943, p. 1452, 1500, Pl. 1, fig. 1; 1953, p. 324–326) photographed dust clouds raised from river silt bars and attributed the phenomenon to strong katabatic winds originating over glaciers in tributary valleys. Péwé (1951, p. 399–401; 1955, p. 720) records this same phenomenon without mentioning earlier observations and uses it as evidence that loess is accumulating in the Fairbanks area today. When viewed by the writer the wind was dissipated on reaching the broad Tanana Valley, and meteorologists who have worked at the U. S. Weather Bureau Station in Fairbanks agree that the dust-bearing winds from Delta River and the Tanana Valley do not reach Fairbanks. Dust in a cabin on the bank of the Tanana is no evidence that loess in now accumulating over an appreciable area, since it probably came from river bars about a mile away. Wind transportation was, possibly, more important at times in the past, but any dust deposited on valley slopes would become mixed with silt of other origin during migration to valley floors, as has happened with volcanic ash.

The valley silts are, therefore, a mixture in varying proportions of materials widely different in origin. The problem is not whether loess or silt formed by weathering is present in the area, but rather: (1) the relative importance of glaciation and frost action in the origin of the silts; and (2) whether they were deposited in their present location chiefly by water or by wind.

Mineral and Chemical Composition

Great differences exist in the mineral and chemical composition of silts in neighboring drainage areas, whereas if they had been transported by wind their composition should be approximately the same. In Goldstream valley silt an exceptionally high muscovite content was noticed in the field, and chemical analysis showed the percentage of K_2O to be 6.1, twice as high as that of any similar silt; but 2 miles away silt from the upper part of Engineer Creek valley, where crossed by the Steese Highway, was relatively low in muscovite, and the K_2O content was only 0.86 per cent. The Engineer Creek silt contained the highest percentage of alumina found in any silt, 20.23; the Goldstream silt contained only 13.9 per cent. This high alumina content is due to the presence of kyanite which was not observed in any other silt.

Péwé (1955, p. 742) reproduces the chemical analysis of the Engineer Creek silt given by Taber (1943, p. 1479–1480), but in spite of the statement that "the percentages . . . are in terms of the ignited material" he adds the

ignition loss to the rest of the percentages to give the absurdly high total of 105.59 per cent.

Petrographic examination shows no evidence of wind sorting with decrease in the abundance of heavy minerals (such as magnetite and ilmenite) with distance from the Tanana although there are high intervening ridges.

Texture

Under the glacio-aeolian theory fine silt is produced by sorting and should become finer with distance from the Tanana, but it does not. In Cleary Creek valley near Chatanika, 18 miles from the Tanana, the percentages of silt finer than 0.074 mm in two samples were 66.3 and 68.0, while near Dome, 12 miles from the Tanana, the percentage was 77.6, and in Ester valley, 4 miles from the Tanana, it was 81.5. The distances and percentages for other localities in this area are: Goldstream, 7 miles and 80.1 per cent; valley slope near Fox, 6 miles and 83.5 per cent; Engineer Creek, 5 miles and 70.0 per cent.

Grains 0.053 to 0.420 mm are customarily classed as fine sand, and 0.420 to 2.0 mm as coarse sand. Péwé's grain-size accumulation curves are not carried far enough to show the percentages of coarse sand except for a sample from Dunbar which contained a little less than 1 per cent. The percentages of coarse sand in samples collected in the area by the writer are: fresh rock flour discharged by Castner Glacier, 0.0; Ester Valley, 0.8; slope mantle southeast of Fox, 0.9; near Donnelly Dome, 1.1; Goldstream Valley, 1.6; slope mantle near Dome, 2.2; Engineer Creek Valley at Steese Highway, 6.9; Cleary Valley (2 samples), 8 and 4.7; slope mantle near Cushman Creek, 46.2. With the exception of the samples from Donnelly Dome and Engineer Creek Valley, coarse sand appears to increase rather than decrease with distance from the source of glacial silt. The relatively high percentage in Engineer Creek Valley may be due to some small granite and granite porphyry intrusions. The absence of coarse sand from the sample of glacial flour may be due to the type of rock over which Castner Glacier moves or to sorting by the stream as it emerges. Analyses of more samples would be helpful since the silts vary in texture vertically as well as arealy.

Fragments of quartz, quartzite, and schist, too heavy to be transported by wind, were found in many silts when large samples were screened although they were not noticed in the field.

Structure

Contrary to Péwé (1955, p. 721), bedding is often well developed in valley silts and is discernible in most good exposures (Taber, 1943, Pls. 7, 8, 11, 13, 14, 16, 17). Silt is interbedded with layers of peat and occasionally thin layers or lenses of sand averaging about an inch thick. An unusually thick bed of sand at the base of silt in Goldstream valley was photographed (Taber, 1943, Pl. 8, fig. 2). At one place pebbly sand, cross-bedded near the bottom, contained thin layers of peaty silt near the top. Interbedding of gravel and silt was observed but is not common.

Since nearly all silt is moved from its place of origin by wind or water, or both, its occurrence on rock of different mineral composition is not in itself evidence of wind transportation, but deposits more than a few feet thick, blanketing high divides, could hardly be formed otherwise than by wind transportation. Such occurrences were not seen by the writer.

Hills of thick bedded silt have been formed by erosion of the surrounding silt. This is evidently the origin of College Hill (Fig. 1) with an elevation of 177 feet above Chena River flood plain, and of Gold Hill which is 127 feet higher. The great difference in grain size of samples from different depths in the drill hole on College Hill indicates bedding and not "vertical uniformity" (Table 1). At depths of 20 and 40 feet the silt contains more than twice as much sand as at 70 feet; at depth of 70 feet the coarse-silt content is nearly 50 per cent higher than at depth of 50 feet; and the fraction with grain size less than 0.01 mm is more than seven times greater at a depth of 50 feet than at 40 feet. These figures are taken from Péwé's (1955, Fig. 8, p. 708) Cumulative-frequency curves of Upland silt from the drill hole near the top of College Hill.

Topography (Fig. 1) suggests double stream capture as an explanation of the origin of College Hill. Under this hypothesis Goldstream Creek after deposition of most of the silt flowed through the low gap at Happy Station into Happy Creek which then flowed on the north side of College Hill. Goldsteam Valley now narrows abruptly northwest of the gap. Capture and diversion of Goldstream into its present course reduced the volume of Happy Creek, thus making possible its capture by a short tributary of Cripple Creek. The complete geomorphic history of the area is, of course, longer and much more complicated. Early workers thought that "Goldstream at

one time flowed to the Tanana by way of St. Patrick [Happy] Creek and that Ester left its valley a short distance below the Eva Creek valley and flowed to the St. Patrick [Happy] Creek" (Prindle and Katz, 1913, p. 105).

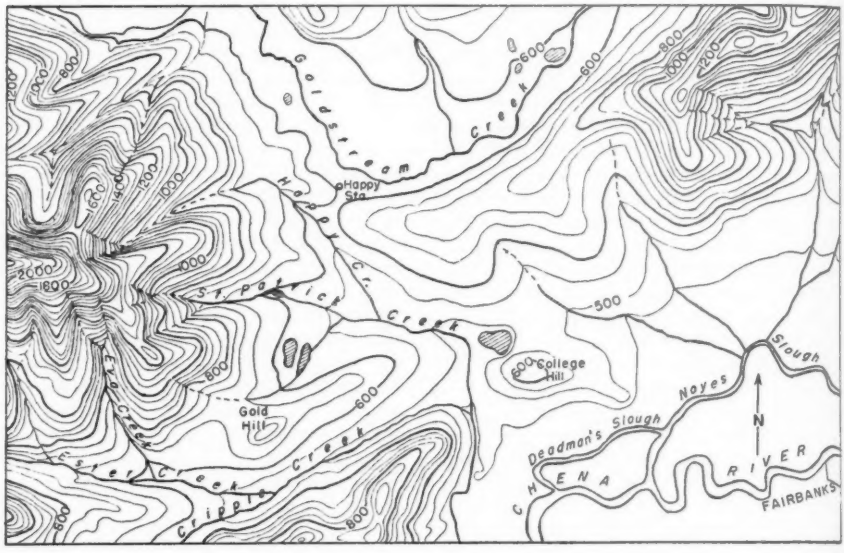

FIGURE 1.—AREA NORTHEAST OF FAIRBANKS, ALASKA
Based on U. S. Geological Survey maps

Gold Hill may be due to stream capture, but the evidence is not so strong as for College Hill.

A model of the spur at the junction of Cripple and Ester Creeks, prepared by the Fairbanks Exploration Co., shows that the two valleys have been eroded to a depth of about 180 feet in silt resting on gravel and occupying a broad bed-rock valley. The gravel, which is slightly lower under the valleys than under the spur, is 160 feet thick on the northwest side of Cripple Creek. Péwé (1955, p. 709) states that this

> "section is composed entirely of uniform- textured silt in spite of slight color variations [bedding?]. Cumulative-frequency curves of mechanical analyses of samples nos. 157, 158, and 159 from this section are strikingly similar (Fig. 9) and also similar to those of silt from the upland (Figs. 6, 7)."

His statement is not supported by his particle-size accumulation curves. They show that the samples have a certain family similarity, since all are silts, but in detail they differ greatly, and these differences are highly significant. The sand fraction is 4 per cent in Sample 157, 16 per cent in Sample 158, and in a sample from Big Delta Air Base (Péwé's Fig. 6) it is 20 per cent, a striking dissimilarity. The other fractions show corresponding differences. A peat layer with fossil tree stump was photographed by Taber (1943, Pl. 9, fig. 2) on the Ester Creek side of the spur.

At Gold Hill about 200 feet of silt rests on auriferous gravel. Bedrock topography indicates the gravel was deposited in a broad valley at the confluence of several creeks. The silt, as elsewhere, rests conformably on the gravel and thins out abruptly upslope. About a mile north of Gold Hill, at the same elevation and extending up the slope of Ester Dome, the overburden formed by rock disintegration and wind deposition is thin enough to permit mapping of bedrock (Prindle and Katz, 1913, Pl. XI) and of mineralized veins (Hill, 1933, Pl. 3). Here as elsewhere thick deposits are

confined to the valleys, and only a thin mantle is found on divides and higher slopes.

The maximum thickness of silt in the College Hill-Ester area must at one time have been well over 200 feet and possibly as much as 300 feet. The great thickness of valley silt removed by erosion was also shown by stripping operations in Goldstream valley near Fox where 150–200 feet of bedded silt resting on the flat surface of underlying gravel was exposed on the southeast side of the valley, but on the northwest side the silt has been almost entirely removed (Taber, 1943, p. 1494, Pl. 6). The upper limit of valley silts in the area shown in Figure 1 is about 800 feet above sea level and at many places is marked by a rather abrupt change in slope. This change in slope appears to have been a major criterion used by Prindle and Katz (1913, Pl. XI) in mapping the Quaternary, since exposures are few. Upstream the valley silt rises, and near the headwaters of Goldstream the elevation is over 1100 feet. This is normal distribution of sediments in a graded stream valley and is evidence against the old lacustrine theory (Spurr, 1898, p. 200–230).

TABLE 1.—DISTRIBUTION OF GRAIN SIZE IN SILT SAMPLES FROM COLLEGE HILL DRILL HOLE

In per cent

Depth in feet	Sand	Silt	
	Greater than 0.05 mm	0.05 mm to 0.01 mm	Less than 0.01 mm
10	10	74	16
20	14	65	21
30	9	70	21
40	13	83	4
50	10	60	30
70	6	88	6
75	8	85	7

On valley slopes above the bedded deposits and on the higher interfluves between the Tanana and Yukon, the silty soil contains numerous angular fragments of the country rock, chiefly schist. They may be seen along the Davidson Ditch of the Fairbanks Exploration Co. and along the Steese Highway between Fairbanks and Circle. The amount of silt deposited by wind in these areas must be very small or the rock fragments would be blanketed.

Age of Valley Silts

Silt accumulation was more important at certain times in the past, but evidence is lacking to support the assumption that the thick valley silts were deposited during periods of glacial expansion. Taber (1943, p. 1481–1491) has shown that most of the creek valley silt accumulated when the climate was slightly warmer and the ground not perennially frozen. A warmer climate is indicated by the distribution of certain fossil plants and animals, and also by ice wedges, extending to depths of as much as 150 feet below the cooling surface, for they had to be formed largely by water from below during the downward freezing of the silt after deposition. Fossil spruce, badly splintered and separated along the annual rings, indicates alteration and saturation by long burial in wet silt prior to freezing.

Radiocarbon dating was unknown when the writer worked in Alaska, but stripping operations in gold mines furnished good exposures of a widespread unconformity marked by deep but not complete thawing of perennially frozen ground, partial removal of thawed silt and redeposition of some of it together with more or less new material. As the climate again became colder, downward freezing of the unfrozen silt occurred. This unconformity was observed on Seward Peninsula as well as in Central Alaska. It is probably the result of more than one period of thawing and erosion, the last during the Thermal maximum some 5000 years ago. It is extremely unlikely that such deep thawing and erosion could have occurred entirely in that relatively short period, and it was not warm enough during the minor retreats of Wisconsin ice for deep thawing in areas now perennially frozen. Therefore, it is tentatively suggested that this thawing took place mostly in one or more of the interglacial stages, possibly chiefly during the Yarmouth, since that interglacial stage was probably the longest and warmest.

The older silt is thicker and more extensive than the younger deposits although a tremendous amount has been removed by erosion. It is too old for C^{14} dating by present methods which are not applicable to samples as old as mid-Wisconsin. Under favorable conditions it contains up to 80 per cent ice by volume, and water to form this ice had to come largely from below, for when a thin layer of perennially frozen ground has formed water cannot penetrate from above. The impermeability of frozen

ground is shown by the presence of artesian water in gravel under frozen silt (Taber, 1943, p. 1520). The younger silts do not contain such large amounts of ice because water could not rise through the frozen ground below. Only the upper part of the older silt shows evidence of having been thawed and refrozen; The bulk of it was frozen after it accumulated and has never been thawed. These facts indicate that the older silt is pre-Wisconsin and may be as old as early Pleistocene. The writer dates the beginning of the Pleistocene from the world-wide crustal movements that occurred shortly before the extreme cold of the first (Nebraskan) ice age and which made possible the intercontinental migration of the large mammals found in the silts and also the underlying gravels.

At many places on flat or gently sloping Alaska tundra, thawing and erosion have left truncated ice wedges forming polygonal networks close to the surface. The shallow trenches above the wedges (Taber, 1943, Pl. 21, fig. 1; Fig. 11) contain only a little slumped silt, whereas, had deposition of wind-borne silt been active during the last few thousand years, a thick deposit of loess should have buried these minor topographic features.

Comparison with Mississippi Valley Loess

Alaska silts have been compared with Mississippi Valley loess, but conditions in the two areas have been very different. The southern ice front extended for many miles with broad outwash plains, whereas glaciers tributary to the Tanana were, during the ice ages, as now, primarily valley glaciers. They did not extend appreciably beyond the base of the mountains which is only 25–50 miles from the crest. Their load of silt was delivered to streams that carried most of it directly to the Tanana. Because of low temperature and light precipitation glacier movement in Alaska was even slower than now, and the glacial mill, therefore, much less efficient than in the south. With change to a warmer climate the ice sheet made an interrupted but relatively rapid northward retreat of 1500 miles leaving large areas unprotected from winds until vegetation could be re-established. Glaciers tributary to the Tanana have retreated only a few tens of miles. In the Mississippi Valley relief is low so that sand and heavy minerals could be swept along on or close to the surface, but for silt to reach Cleary Valley from the Tanana it would have to be carried for 25 miles across a maturely dissected upland with valleys 2000 feet below the ridges. The winds as recorded at Fairbanks are remarkably weak, averaging 5.2 miles per hour, half the speed of wind at St. Louis, and a third the speed at Chicago. During the ice ages the difference was probably much greater.

Conclusions

An examination of available evidence indicates that: (1) although glaciers have produced much silt in limited areas they could not have produced the tremendous amount present in the Fairbanks district plus the large amount that has been removed by erosion, (2) most of the silt has been formed from Birch Creek Schist by frost disintegration, and (3) the thick valley silts, which form the bulk of the deposits, have been deposited in their present location by water.

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